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ESTIMATES OF FLIGHT REGIMES IN THE VENUS ATMOSPHERE

Peter P. Wegener

The RAND *Corporation*

SANTA MONICA • CALIFORNIA

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PREFACE

The purpose of this study was to gain an insight into the aerodynamic flight regimes that may be encountered during entry into the atmosphere of Venus, and to estimate the aerodynamic behavior of objects entering that atmosphere. This in turn led to estimates of the atmospheric composition itself; these estimates were essentially completed in May of 1961, before the Mariner II flyby. Another model of the Venus atmosphere, based on some recent interpretations of observed data, is being included in RM-3388-PR, **Flight Regimes in the Atmospheres of Mars and Venus**, by the same author, a RAND consultant.

SUMMARY

This is a brief elementary review of some of the methods employed in the determination of composition and structure of planetary atmospheres. The atmosphere of Venus is discussed on the basis of existing evidence. In view of the possible aerodynamic effects of the atmosphere of Venus, the findings reported in the literature are categorized conservatively as either accepted or uncertain.

Two limiting atmospheric models and some variants are constructed with the intent of encompassing the range of uncertainty of responsible opinion. Free-flight Mach numbers and Reynolds numbers as a function of altitude are computed for these models, and a simple entry trajectory and satellite flight conditions are discussed in terms of these models.

It is found that the aerodynamic problems of the troposphere are very different for the limiting models. It is further seen that the continuum flight regime is entered at altitudes between about 150 and 80 km above the visible cloud layer, depending on the models used. Finally, chemical kinetics problems, heat transfer, etc., are expected to show major variations, and further work in this area is suggested.

ACKNOWLEDGMENTS

CONTENTS

LIST OF SYMBOLS

a = low-frequency sound speed

A = Angstrom unit

a.u. = astronomical unit, distance from sun to Earth, p. 9

c = mean molecular speed

g = acceleration due to gravity

h = altitude (also Planck's constant)

H_o = scale height, defined by Eq. (6)

K = Knudsen number

l = characteristic length

m = polytropic exponent

M = Mach number

p = pressure

$\dot{q}$ = heat-transfer rate

r = radius of planet

R = universal gas constant

Re = Reynolds number

t = time

T = absolute temperature

u = flight speed

v = specific volume

x = mole fraction

β = dry adiabatic lapse rate, defined by Eq. (17), or temperature gradient

γ = ratio of the specific heats

λ = mean free path of the molecules

ν = frequency

ρ = density

η = absolute viscosity

μ = molecular weight

Subscripts

i = i[th] species

c = Cytherean visible cloud layer

s = planetary surface

∞ = free-stream conditions (or incident undisturbed flow)

o = standard pressure and temperature (STP)

Conversion factors

1 A = 10^{-8} cm

1 atmosphere = 760 mm Hg = 1.013 bar

1 bar = 750.1 mm Hg = 10^{6} dyne/cm^2

1 mb = 1 millibar = 10^{-3} bar

1 micron = 10^{-3} mm

1 poise = 1 g/cm sec (unit of viscosity)

R = 8.31 x 10^{7} erg/$^{\circ}$K mole = 82.06 cm^3 atm/$^{\circ}$K mole

μ_{air} = 28.97

μ_{N_2} = 28.02

μ_{CO_2} = 44.01

I. ELEMENTS OF THE STUDY OF PLANETARY ATMOSPHERES

From the time of the origin of the solar system, various physical and chemical processes have produced gases that surround planetary surfaces. Although in principle these processes are similar, the current composition and structure of planetary atmospheres are different. This difference depends on the position of the planet in the solar system, i.e., its distance from the Sun, as well as its mass and the detail of the chemical reactions on the surface and in the interior. A planet's energy balance, which differs with distance from the Sun, is established by the rates at which energy is received from the Sun and radiated from the planet. Differences in gravitational attraction have acted selectively on gases with different molecular weights. In fact, small planets like Mercury (or the Earth's Moon) have lost nearly all gases to interplanetary space. From our knowledge of the Earth, three major processes are recognized in the development of the atmospheres of the planets:[1,2] gases escape if molecules reach escape velocity, which is estimated from kinetic theory; gases are produced in the planet's interior and diffuse to its surface; and finally, gases are produced by chemical reactions of the materials composing the planet's surface. The current status of planetary atmospheres may therefore be inferred to some extent from considerations of a geochemical or morphological nature.[1,2,3] The danger inherent in this approach to determining the current nature of planetary atmospheres is our lack of ability to consider with sufficient accuracy the complicated interplay of chemical reactions taking place in the geological time scale. Therefore many astronomers appear to believe that only direct observation ought to be relied on to estimate the current composition and structure of planetary atmospheres.

The observational study of planetary atmospheres rests on an interpretation of the radiation received from the planet. This radiation has two origins: it stems mainly from the reflected and scattered sunlight, and secondarily from radiation emitted at some height in the planet's atmosphere (or from its surface). The restriction on radiation studies of reflected sunlight from the viewpoint of an earthbound

observer is the fact that the radiation passes through the Earth's
atmosphere in addition to passing twice through the planetary atmos-
phere under study. Furthermore, the double light path through the
planet's atmosphere is longer near the "edge" or terminator of the
planet than at the midpoint of the equator. The Earth's atmosphere,
because of its own composition, is transparent to radiation only in
certain regions of the spectrum, and these regions are called "windows."
Other parts of the spectrum are shut out because of the complete ab-
sorption of radiation by the constituents of our own atmosphere. A
transmission spectrum of the Earth's atmosphere has windows in the
infrared, the near-infrared, the visible, and the ultraviolet.[*] In
this range of wavelengths, observations are made of sunlight reflected
from the planet. Such observations may be made visually with tele-
scopes, or by taking photographs at various wavelengths, measuring
thermocouple temperatures through telescopes, taking spectra in certain
ranges of wavelength, observing the polarization of the reflected sun-
light, etc. In addition, emission of radiation in the microwave re-
gions (wavelengths on the order of 1 cm) may be observed, and the
origin of this radiation may be ascribed to some layer in the plane-
tary atmosphere or to the surface of the planet, depending on its esti-
mated causes. The last major optical technique concerns the trans-
mission and refraction of light originating from a fixed star during
its occultation[**] (or near-occultation) by a planet. The fixed star's
light can be observed passing through a planetary atmosphere, and with
the actual position of the star in relation to the planet being well
known, the observed bending and absorption of the star's light give
clues to the density, density gradient, etc., of the planet's atmos-
phere.

All the above methods have, at one time or another, been employed
by many observers, and they have yielded a variety of qualitative and
quantitative results. However, many astronomers have unfortunately

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[*] For example, see Fig. 2 of Ref. 4.

[**] Shutting off of the light of one celestial body by the inter-
vention of another.

turned away from planetology in recent years, although much could be gained by repeating previous work with the improved, modern, earth-bound instrumentation now available. Furthermore, a new tool has recently been added that requires extensive application: The first observations of planetary atmospheres have been made from high-flying balloons operating above the major portion of our own atmosphere. Also, there are exciting prospects for the near future, when such observations can be improved by the use of orbiting observatories.

Most quantitative results on the composition of planetary atmospheres are obtained by taking spectra of infrared radiation, a method by which, in principle, molecular species present may be identified, and the total amounts present in a planetary atmosphere may be measured. The equipment required for these experiments is highly refined and especially built for the purpose,[*] and the observers must be experienced. Such observations must be viewed together with comparison spectra, usually taken of the Moon, whose atmosphere may be considered negligible in this context. The two spectra are then compared, and in this manner the telluric[**] lines and bands may be disentangled from those caused by absorption of unknown species in the planet's atmosphere.

An aid in the separation of spectral lines of the planet's atmosphere from those caused by our own atmosphere is provided by the Doppler effect: As the planet approaches or recedes from the Earth, a given line produced by a constituent in the planet's atmosphere and the corresponding line in the Earth's spectrum will be separated, with the planet's line shifting to the blue (higher-frequency) or red (lower-frequency) part of the spectrum, respectively. To the untrained observer, however, the planetary and comparison spectra look much alike,[***] with the planet adding a few "wrinkles." Only very careful photometry will reveal the quantitative difference between the spectra.

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[*] For example, see p. 288 of Ref. 3.

[**] Pertaining to the Earth, terrestrial.

[***] For example, see Mars and Moon spectra by Kuiper, p. 359 of Ref. 3.

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The final problem in the interpretation of the spectra obtained lies in the fact that the absorption in the planetary atmosphere takes place along a long path at relatively low pressure above a certain re- flection layer, where the direction of the light path is reversed for the wavelength in question. Only limited information on absorption by the gases under discussion (say CO_2 on Venus), taken under similar con- ditions in the laboratory, is available. In fact, before such labora- tory experiments were made, the CO_2 absorption bands seen on Venus, for example, had never been observed. Once a constituent has been identified, temperature estimates may also be made from the spectrum. The rotational degree of freedom of the molecules leads to a fine struc- ture of the spectrum and equivalent black-body temperatures may be com- puted. (However, such a result on temperature cannot readily be ascribed to a given height.)

Visual observation (and photography) is a somewhat secondary source of information and is usually qualitative only. The presence of an atmosphere may be noted generally (as it was first by Schroeter in 1796 for Venus[5]) by extension of the crescent of the partly illumi- nated disk beyond a semicircle. For example, variable markings in the seemingly permanent cloud cover of Venus have been observed (photo- graphy in the ultraviolet shows these quite clearly[6,7]), and haze has been seen on Mars. However, little quantitative deduction has been possible.

Collecting evidence for the presence of certain constituents and estimating temperature (or pressure) at certain heights by these methods will lead to a variety of results, which must finally be pieced together to form a consistent picture of a planetary atmosphere. Certain temper- ature determinations may be ascribed to a certain interval of altitude. Amounts of gases deduced from spectra must be present above a height that is taken to be the effective reflecting layer for the particular wavelength in question. In this manner, a partial pressure of a gas may be estimated. This reflecting layer may be the surface, a "cloud layer," or some other layer at a given altitude. With certain tempera- tures, partial pressures, etc., available at certain heights, and with the assumption of a certain composition, the structure of the atmosphere

may be estimated. Such calculations are based on the integration of the basic equation of hydrostatics, with certain assumptions on variables in a given height interval. In one instance, the temperature and molecular weight of the mixture may be assumed to be constant in the interval, which leads to the well-known exponential atmosphere. In other cases a temperature distribution may be given, or at least two temperatures at two heights, and the pressure and density distribution may again be estimated. The variation of composition and therefore of molecular weight, important at high altitudes, may also be included. In the composite picture of the atmosphere from the planet's surface to the exosphere, certain compatibility conditions must be met to insure that the atmosphere remains stable. Such calculations are discussed in textbooks on meteorology and some of them (from the fluid-dynamics viewpoint) are outlined in Appendix A for use with the results of this Memorandum.

The total of all available information on planetary atmospheres does not meet in any sense the stringent requirements demanded by the aerodynamicist, who wishes to compute trajectories, aerodynamic characteristics, etc., in order to design a vehicle that may penetrate the atmosphere, maneuver, and land in an unburnt condition. There is major disagreement among different authorities on the subject of the atmospheres of even our nearest planets, and it is well to keep in mind that many results are in flux and are still in the nature of speculations. However, faced with this situation, it may be permissible to construct limiting models of planetary atmospheres, working with limits encompassing the range of responsible opinion. At no time, however, should these models be considered more than rough estimates to indicate limiting flight conditions.

Further information on the methods by which planetary atmospheres are studied may be obtained in semipopular and technical books.[*] Concerning the planet Venus in particular, the instructive article by Sagan in Ref. 6 is recommended.

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[*]See Refs. 1, 2, 3, 5, 8, 9, 10.

II. THE ATMOSPHERE OF VENUS

Some of the elements of the orbit of Venus and some of the physical characteristics of the planet are given below.[*]

Orbital Elements

Semimajor axis = 0.723 a.u.[**] = 1.082 x 10^8 km

Period of revolution around Sun = 224.7 days = 0.615 year

Mean orbital velocity = 35.05 km/sec

Eccentricity = 0.0068

Inclination to ecliptic = $3°23.6'$

Physical Characteristics

Equatorial diameter[***] = 12,400 km = 0.97 (Earth = 1)

Volume = 0.91 (Earth = 1)

Mass = 0.815 (Earth = 1)

Surface gravity = 860 cm/sec^2 (mean value)[11]

Escape velocity = 10.3 km/sec

Period of rotation = less than period of revolution, more than one day, probably on the order of one week.

Albedo[****] = 0.59[2] or 0.76[3] (highest in solar system)

Solar radiation received outside the atmosphere = 1.90 (Earth = 1), computed from relative distances to the Sun.[*****]

Solar constant at surface[2] = 1.3 (Earth = 1) for an assumed albedo = 0.59.

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[*]See p. 413 of Ref. 2, p. 398 of Ref. 4, and p. 95 of Ref. 10.

[**]One astronomical unit (the distance from the Sun to the Earth) is approximately 1.50 x 10^8 km.

[***]Refers to the top of the visible cloud layer--approximate.

[****]Ratio of reflected to incident light.

[*****]Solar constant (Earth) = 1.395 x 10^6 ergs/cm^2 sec.

GENERALLY ACCEPTED PROPERTIES

The division of the findings into generally accepted and uncertain properties of the Cytherean[*] atmosphere, to be discussed next, differs to some extent from that proposed by the astronomers. The viewpoint in this study has primarily been guided by the aerodynamic consequences that would result from accepting or not accepting a certain result or speculation. Such doubts will lead to the limiting atmospheric models whose effects on aerodynamics, in particular at low altitudes, will indeed be marked. (In this discussion reference will often be made to the secondary literature only.)

The following gives a list of suggested constituents of the Venus atmosphere:

Substance	Remarks
CO_2	Refs. 2, 3, 12, spectroscopic evidence.
CO	Refs. 2 and 3, spectroscopic evidence.
N_2O, CH_4, C_2H, C_2H_6, NH_3	Ref. 3, spectroscopic evidence.
H_2O	Strong, Ref. 9 of Ref. 6, balloon observation.
N_2	Ref. 2, ion bands in the night sky; also analogy to the Earth.
A	H. Brown, p. 258 in Ref. 3, geochemical deduction.
O_2 (?)	De Vaucouleurs, p. 55 in Ref. 10.
Other minor constituents	Ref. 6; also Kopal (who disagrees in Ref. 13 with line 3 in this table).

Of the constituents listed, large amounts of CO_2 and N_2 only are expected. The reason for assuming the presence of N_2 as an "inert filler" gas (as on Mars) is primarily based on the facts known for the Earth, since N_2 is spectroscopically inaccessible to us. Most other constituents (with the possible exception of argon[**]) are present in

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[*] Of or pertaining to the planet Venus. There is an amusing discussion of this word by Sagan in Ref. 6.

[**] See Brown on p. 266 of Ref. 3.

very small amounts only. From the aerodynamic viewpoint the minor constituents may generally be neglected in view of the other uncertainties. However, their possible later role in the detail of chemical kinetics problems in shock waves, heat transfer, etc., must be considered. Water vapor in small amounts, actually discovered by Strong[6] (20 microns*), probably furnishes the final clue to the composition of the clouds. Oxygen has not yet been discovered, but a small amount is postulated by de Vaucouleurs. At any rate, it is expected that the dissociation of CO_2 by sunlight produces free O, and from the reaction $O + O + M \rightleftarrows O_2 + M$, where M may be any third body, some O_2 will be continuously produced. This O_2 will again be dissociated into $O + O$, but in equilibrium some O_2 should always be present. However, the aerodynamicist's problem lies primarily in the interpretation of the quantitative amounts of CO_2 and N_2, and unfortunately this question must be relegated to the section on uncertainties.

The surface of Venus has never been seen, since the planet is shrouded in visually impenetrable cloud cover of unknown composition. All radius estimates, etc., are approximate, and they are based on the measurement of the top of this visible cloud layer. Atmospheric-structure calculations will have to be normalized to this altitude.**

Of the temperatures measured, it appears established that thermocouple measurements made in the infrared[14,15] (which show little difference between the day and night side) give a temperature of about 235°K. This temperature is probably close to that of the top of the cloud layer.[2,6] Chamberlain and Kuiper[16] derived a temperature of about 285°K from the distribution of intensities in the fine structure of the rotational part of the spectrum. This higher temperature may be taken as "characteristic of the average temperature above the

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*This number implies that if all water vapor present were precipitated as a liquid at 1 atmosphere of pressure, the liquid layer would be 20 microns thick.

**It is more precise to speak of a "reflecting layer" or "reversal layer" for the radiation in question--in this case the 8000-A bands on the basis of which most of the following deductions are made. Other reflecting layers exist at different altitudes.

clouds";[2] it cannot be assigned to a particular altitude. From several other observations and deductions[1,2,3] it appeared reasonable, until the advent of microwave measurements, to assume an average surface temperature for Venus of about $350°K$. Microwave measurements raised this value to about $600°K$, however, and the aerodynamic consequences of accepting this higher value are so drastic that in view of other existing uncertainties, these findings will be relegated to the next section. Finally, the temperature must be expected to rise greatly at higher altitudes in the ionosphere of Venus, which begins about 80 to 100 km above the cloud layer, where dissociation of CO_2 comes into play. This rise is caused by chemical reactions under the influence of sunlight ($h\nu$), and Sagan[6] lists the following processes:

$$O + h\nu = O^+ + e^-$$

$$CO + h\nu = CO^+ + e^-$$

$$CO_2 + h\nu = CO + O$$
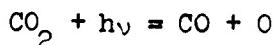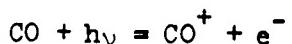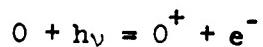
The actual temperature gradient may at best be roughly estimated.

In general terms it may safely be assumed that Venus has a troposphere (lower atmosphere), as does the Earth, with a temperature that decreases to close to dry adiabatically (?) with altitude the tropopause (upper edge of the troposphere) is likely to be located at or near the top of the visible cloud layer (or reflecting layer in the 8000-A bands). A stratosphere exists which may be very nearly isothermal. And finally, an ionosphere is present, beginning about 80 to 100 km above the cloud layer. Temperatures in the ionosphere increase because of the reactions discussed. It is accepted, regardless of the actual CO_2 concentration, that Venus has a strong greenhouse effect.* This effect occurs if most of the solar radiation reaches the surface of the planet while little of the heat of the surface is lost to interplanetary space.

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*Sunlight, whose energy maximum lies in the visible range of the spectrum, penetrates the glass roof of a greenhouse. The energy is converted to heat in the soil, and since glass does not pass the infrared emission, little heat escapes to the outside. On Venus, CO_2 and H_2O take the role of the glass.

The recent occultation of Regulus, as interpreted by Kaplan,[17] led de Vaucouleurs and Menzel[18] to state a pressure of 2.6×10^{-3} mb at a height of 70 ± 8 km above the visible cloud layer.

Unfortunately, it appears that the recitation of established facts must be terminated at this point. In particular, it appears to be difficult to quote further values of pressure as a function of altitude because these values result from assumptions about the quantitative distribution of the major constituents.

AREAS OF UNCERTAINTY

The work on near-infrared spectra of Venus done by Kuiper and his coworkers* provides the basis for the estimates of the amount of CO_2 present in the atmosphere of Venus. This work has been interpreted with the aid of laboratory experiments in the far red and the infrared by Herzberg and Kuiper.** In these experiments, spectra of CO_2 were taken in long tubes with multiple radiation paths in order to approximate the conditions of absorption in a planetary atmosphere. Similar experiments were carried out more recently by Howard et al.[19] Basing their work on Herzberg's experiments, he and a number of other authors[2,3] find that approximately 1 km (at standard pressure and temperature) of CO_2 is present in the Cytherean atmosphere above the effective reflecting layer corresponding roughly to the 8000-A bands. On this basis it is possible to assign a partial pressure of about 0.17 atmosphere to that layer above the surface where the radiation is reflected (or the path is reversed). The method of this calculation is described in Appendix A. Neglecting minor constituents, it is reasonable to assume that in addition to the CO_2, about 10 mole per cent of N_2 is present as suggested by Urey.[2]

It is interesting to note that Dole[20] arrives at the same composition of the Cytherean atmosphere (90 per cent CO_2, 10 per cent N_2)

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*See p. 306 of Ref. 3 and Refs. 5 to 7 of Ref. 17.

**See p. 406 of Ref. 3.

from geochemical deductions alone. He applied to Venus the current thinking on the evolution of our own atmosphere. However, Dole's figures for surface pressures are higher than would be expected from the model above in conjunction with calculations of structure. On the other hand, a new evaluation of Kuiper's original spectra in conjunction with the recent laboratory work of Howard et al.[19] led Kaplan[17] to postulate the Cytherean constituents in a different manner. Kaplan suggests essentially a reversed composition, leading to only about 15 mole per cent of CO_2, with nitrogen as the major component. In turn, the pressure at the reflecting layer would only be 0.033 atmosphere, or about 1/5 the previously assumed pressure.

The second major uncertainty is connected with the observations of microwave emission from Venus. Such observations have been made by several authors; a summary and interpretation of their results is given by Barrett,[21] and the results have also been critically discussed by Sagan.[6,11] Barrett constructs a model Cytherean atmosphere in order to evaluate the absorption properties for radiation of wavelengths from about 1 to 10 cm. The suggestion is offered that the microwave radiation received from Venus is thermal and that it originates at the surface. Equivalent black-body temperatures (or brightness temperatures) may be computed, and these estimated values are tabulated by Barrett[21] and plotted by Sagan.* A reasonable mean value of all the results appears to be $580^{\circ}K$, and this temperature may be assigned to the surface. Sagan[11] investigates in detail the processes which would be required in the ionosphere** of Venus to explain these results as emission of this radiation at high altitude. He finds no likely explanation of microwave emission in the ionosphere, but he does find that the rapid drop-off of brightness temperatures for wavelengths smaller than 1 cm is indeed suggestive of the thermal origin at the surface. Sagan concludes that the far-infrared absorption of

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*See Fig. 1 of Ref. 11.

**Synchrotron radiation from charged particles trapped in a Cytherean Van Allen belt.

energy in the Cytherean atmosphere is sufficient to explain the strong
greenhouse effect required for such a high surface temperature. How-
ever, this explanation also requires the presence of water vapor in
the atmosphere to aid in absorption, as well as an appreciable atmos-
pheric circulation. Such circulation would require Venus to rotate
nonsynchronously, i.e., with a period that is much higher than the
period of revolution. Water vapor has in fact been observed since then
by Strong,[*] and several arguments speak for a period of rotation on
the order of one week.

However, the consequences of accepting the microwave result as
firmly established are indeed grave. A surface temperature of about
$350^\circ K$, which is below the boiling point of water at atmospheric pres-
sure, indicates that Venus could be covered by "vast oceans,"[2] where-
as accepting a surface temperature of $600^\circ K$ turns Venus into an arid
desert-like planet, a picture also suggested by Dole.[20] For a sur-
face temperature of $350^\circ K$, the troposphere would extend to about 10 km
from the surface, and the surface pressure would be about that of Earth.
For $600^\circ K$, the troposphere would extend from 30 to 40 km above the
surface and the surface pressure would be 5 or more atmospheres. (Sur-
face pressures up to 30 atmospheres are compatible with the microwave
emission from the surface, according to Barrett.) The aerodynamic
problems for both cases would be very different, with the high-surface-
temperature model leading to extreme Reynolds numbers of flight. It
therefore appears that a conservative view ought to consider both
possibilities at present. We have no knowledge of the magnetic field
of Venus or of the electron density in the ionosphere to rule out defi-
nitely the possibility that the microwave radiation is emitted at a
high altitude. If Kaplan's thoughts prove to be accurate, we may find
that the greenhouse effect is not potent enough to sustain a surface
temperature of $600^\circ K$. It is for these reasons that the author rele-
gates this finding to the category of uncertainties.

The final uncertainty pertains to the nature of the visible clouds.
Historically, many explanations have been submitted. Lyot's results

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[*]See Ref. 9 of Ref. 6.

on the polarization of reflected sunlight[*] have been superimposed on polarization measurements of liquid water droplets and some agreement is found. Menzel and Whipple[22] suggested, primarily as a result of Lyot's work, that the clouds are water droplets. However, in view of the extreme difficulties of the interpretation of light-scattering data, as pointed out by van de Hulst,[**] and considering the arguments of Dollfuss,[23] Lyot's and Dollfuss's measurements would be equally plausible if the clouds were made up of dust particles. The recent actual observations on the presence of water vapor led Sagan[11] to conclude that the clouds are in fact ice crystals. Combining possible temperature distributions, the author also finds that saturation is possible at the visible cloud level, but higher H_2O partial pressures than those postulated by Sagan are required. However, only slight shifts in the temperature distribution or water-vapor content are sufficient to satisfy a condition of sublimation of ice, and it is therefore likely that the clouds are ice. In each case, these findings are not affected by accepting either $350^\circ K$ or $580^\circ K$ as surface temperature. Aerodynamically speaking, this will be a minor point.

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[*] See p. 386 of Ref. 2.

[**] See p. 49 of Ref. 3.

III. LIMITING MODELS OF THE ATMOSPHERE OF VENUS FOR AERODYNAMIC STUDIES

Traditionally, the first view taken by the aerodynamicist of flight regimes concerns the motion of a body at a given altitude, in terms of free-stream Mach number and free-stream Reynolds number. These parameters may readily be determined for known flight speeds and for known atmospheric compositions and structures. The boundaries of the flight regimes of free-molecule flow, transitional flow, and continuum flow may thus be estimated. Minor constituents of the atmosphere have minor effects on free-stream conditions, since only slight changes of the molecular weight, viscosity, etc., of the gas mixture are involved.

The second and more detailed view considers the properties of the flow in the vicinity of the body, the shock layer, the boundary layers, and their complicated interaction. Minor constituents may now play a major role in high-speed flow. High-speed flow may be defined in this context as any flow in which the thermodynamic state of the gas mixture in the vicinity of the body is such that a thermally and calorically perfect gas may no longer be assumed. Excitation of the vibrational levels of diatomic or polyatomic molecules, dissociation, chemical re-actions, ionization, and electronic excitation lead to departures from the perfect-gas laws; the details of the flow are indeed influenced by small components present in the planetary atmosphere, particularly if relaxation processes of the various modes enter. A typical example pertaining to Venus would be the strong effect exerted by a very small amount of water vapor on the relaxation time of the vibrational modes of CO_2.[24] The free-stream condition will be considered first in this Memorandum. The resulting flight conditions may then form the basis for later calculations of flow fields near the body,[25] with the effects of minor constituents being added.

One of the limiting aerodynamic models of the Cytherean atmosphere will therefore contain pure CO_2, and the other will have N_2 as the ma-jor component. Variants of these models will be discussed by truncating the atmosphere structurally, i.e., by assuming different temperature distributions for fixed composition above and below the visible cloud

level. This will lead to drastic differences for the troposphere, depending on whether the high or the low surface temperature discussed in the last section is selected. The presence of argon has been omitted in all models.

THE CO_2 MODEL

For one limit, it will be assumed that the Cytherean atmosphere contains CO_2 only, 1 km of which exists above the tropopause at normal temperature and pressure. (References discussed in the previous section will not be repeated.) Two variants according to structure will be distinguished. The simplest one, CO_2 Model I (see page 46 of Appendix B) shows a dry adiabatic property distribution from the surface to the tropopause. An isothermal atmosphere at the rotational temperature of $285°K$ is assumed above the cloud level, leading to an exponential decrease in pressure and density. The surface temperature is based on the microwave results, and to be conservative, $T_s = 600°K$ has been chosen. The ratios of the specific heats given above and below the cloud level are compatible with thermodynamic data at the temperatures and pressures involved.[26] Carbon dioxide is not dissociated under surface conditions or in the upper atmosphere[*] and ionization has been neglected. The resulting structure is shown in Figs. 1 and 2. Also shown in Fig. 2 is a pressure distribution, including lower pressure ratios computed for Barrett's model.[21] Barrett postulated an atmosphere of 80 per cent CO_2 and 20 per cent N_2 (with varying amounts of H_2O) in order to evaluate the microwave data. He assumed a surface temperature of $580°K$, and it is this fact rather than the presence of N_2 which leads to lower pressure ratios as a function of altitude in the troposphere.[**]

The CO_2 Model II (see page 47 of Appendix B) is based on the lower surface temperature of $350°K$, and Urey's cloud level or tropopause temperature of $235°K$ (Fig. 1) is assumed. An isothermal stratosphere extends 85 km above the cloud level, and the rise of temperatures at higher

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[*] See Figs. 2b and 2c of Ref. 27. .

[**] See Fig. 3 of Ref. 10.
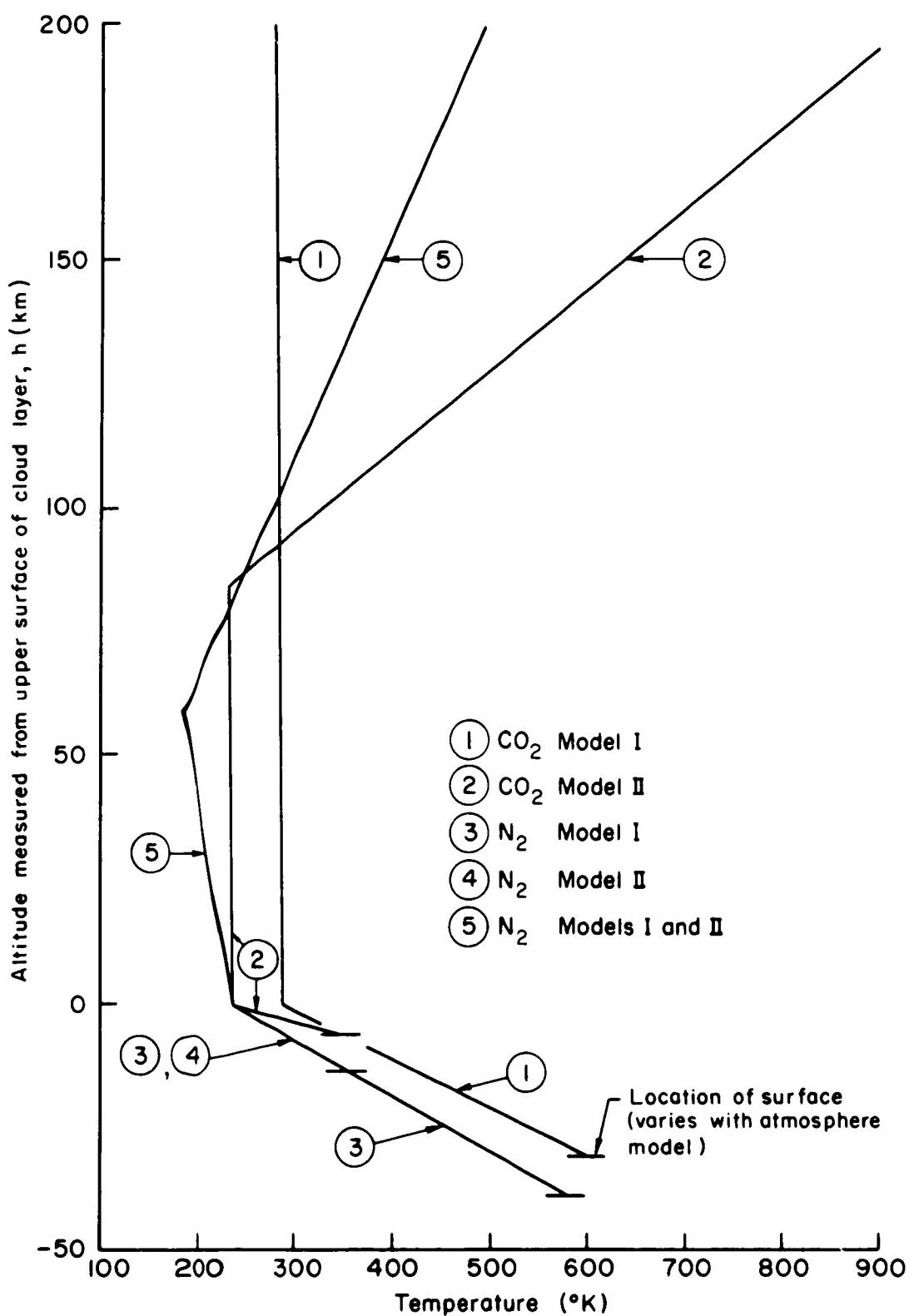
Fig. I—Temperature distribution of models
of the Venus atmosphere
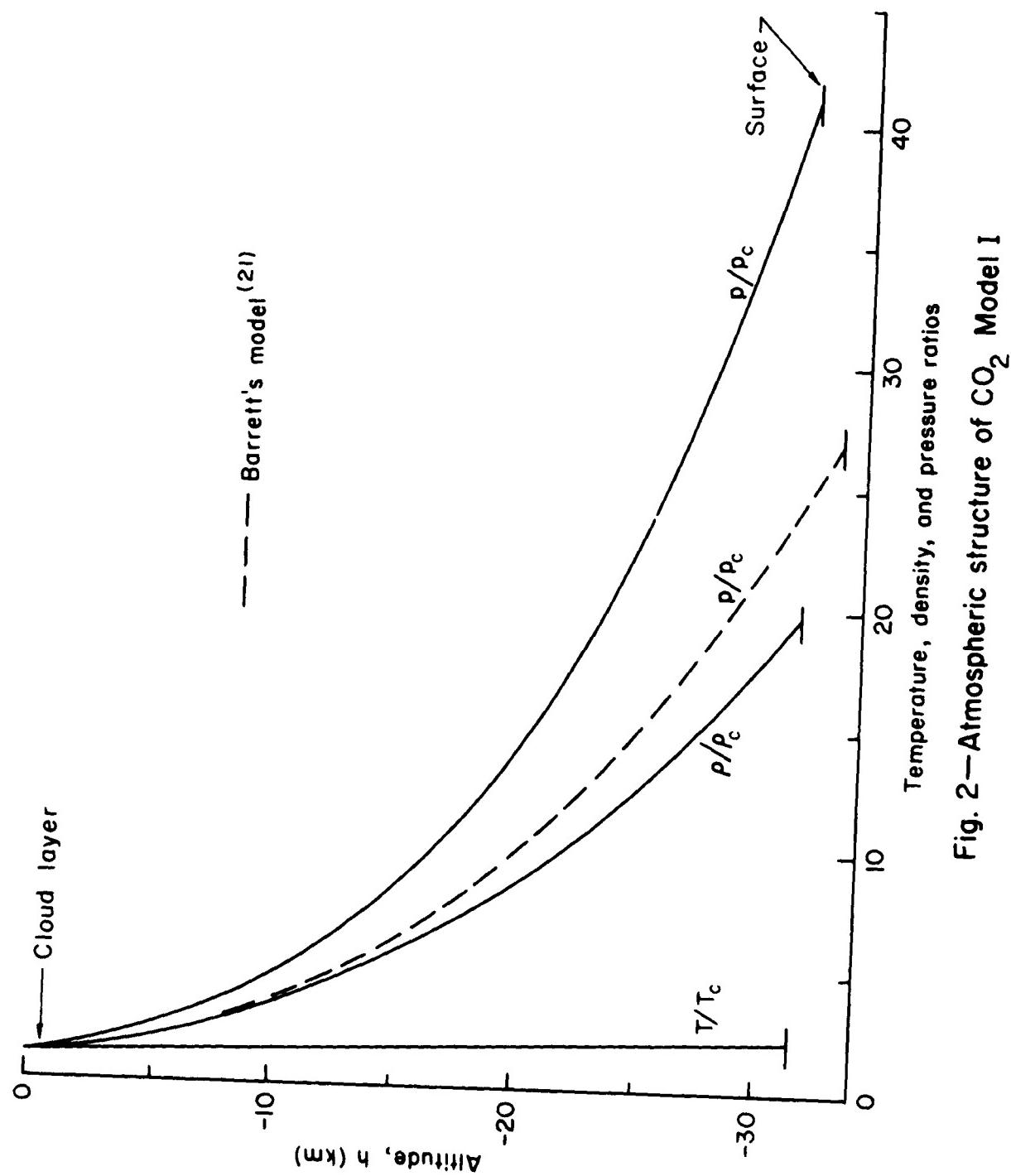
Fig. 2—Atmospheric structure of CO_2 Model I

altitudes follows Urey and Sagan.[*] This temperature gradient represents nothing more than a rough guess. The properties of this variant of the CO_2 Model II for the troposphere are shown in Fig. 3; the result shows a striking difference from CO_2 Model I. The upper atmosphere conditions are shown in Fig. 4 for CO_2 Models I and II, the simple exponential pressure and density distribution of CO_2 Model I being in agreement with the results on the occultation of Regulus. (The mean free path is given in Fig. 8.)[**]

In CO_2 Model II, dissociation (and also ionization reactions) of CO_2 at the highest altitudes is neglected, i.e., the composition and the molecular weight are left unchanged. At $950°K$ at 200 km above the cloud level, there is no dissociation when $p = 10^{-4}$ atm; however, in fact $p = 10^{-11}$ atm,[***] and some dissociation is to be expected.[****] Since the mean free path will be on the order of 10^7 cm (Fig. 8), aerodynamic forces will be negligible, and our lack of reliable information does not permit even a rough guess at the actual conditions.

THE N_2 MODEL

The other limit of our aerodynamic models of the Cytherean atmosphere follows the thoughts on composition predicted by Kaplan.[17] It is assumed that CO_2 provides only 15 mole per cent of the gas mixture, with N_2 the major component at 85 mole per cent. The results for the N_2 model are shown on page 48 of Appendix B. Again the structure below the tropopause includes two choices: one variant, N_2 Model I, is based on the high microwave surface temperature, this time taken as the mean value[6,21] of $580°K$ (Figs. 1 and 5); the other variant, N_2 Model II,

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[*] See Fig. 2 of Ref. 2 and Fig. 3 of Ref. 6.

[**] For comparison, the mean free path for Earth is also shown on Fig. 8 using values taken from the ICAO Standard Atmosphere. The Earth's tropopause was taken as $h_c = 11$ km.

[***] See Fig. 2g of Ref. 27.

[****] The increasing temperature at high altitudes is due in part to direct heating by sunlight and in part to heating from the released dissociation energy. The assumed temperature distribution is based on both processes.
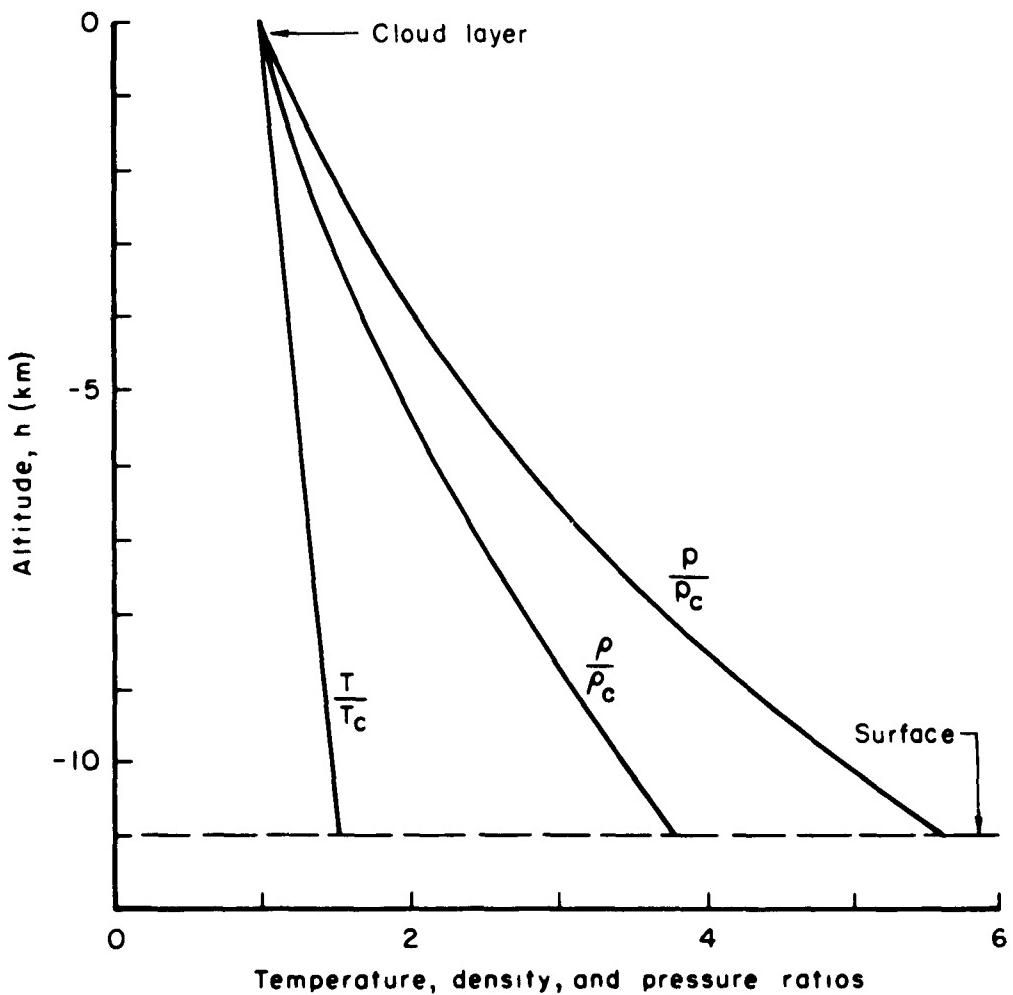
Fig. 3 — Properties of CO$_2$ Model II troposphere

n_2

$n_1 = \log p/p_c = \log \rho/\rho_c ,\, I$

$n_2 = \log p/p_c ,\, II$

$n_3 = \log \rho/\rho_c ,\, II$

n_3

n_1

Occultation-of-
Regulus result
$\left(\log p/p_c\right)$

Altitude above cloud layer, h (km)

200

150

100

50

0

-15   -10   -5   0

log n
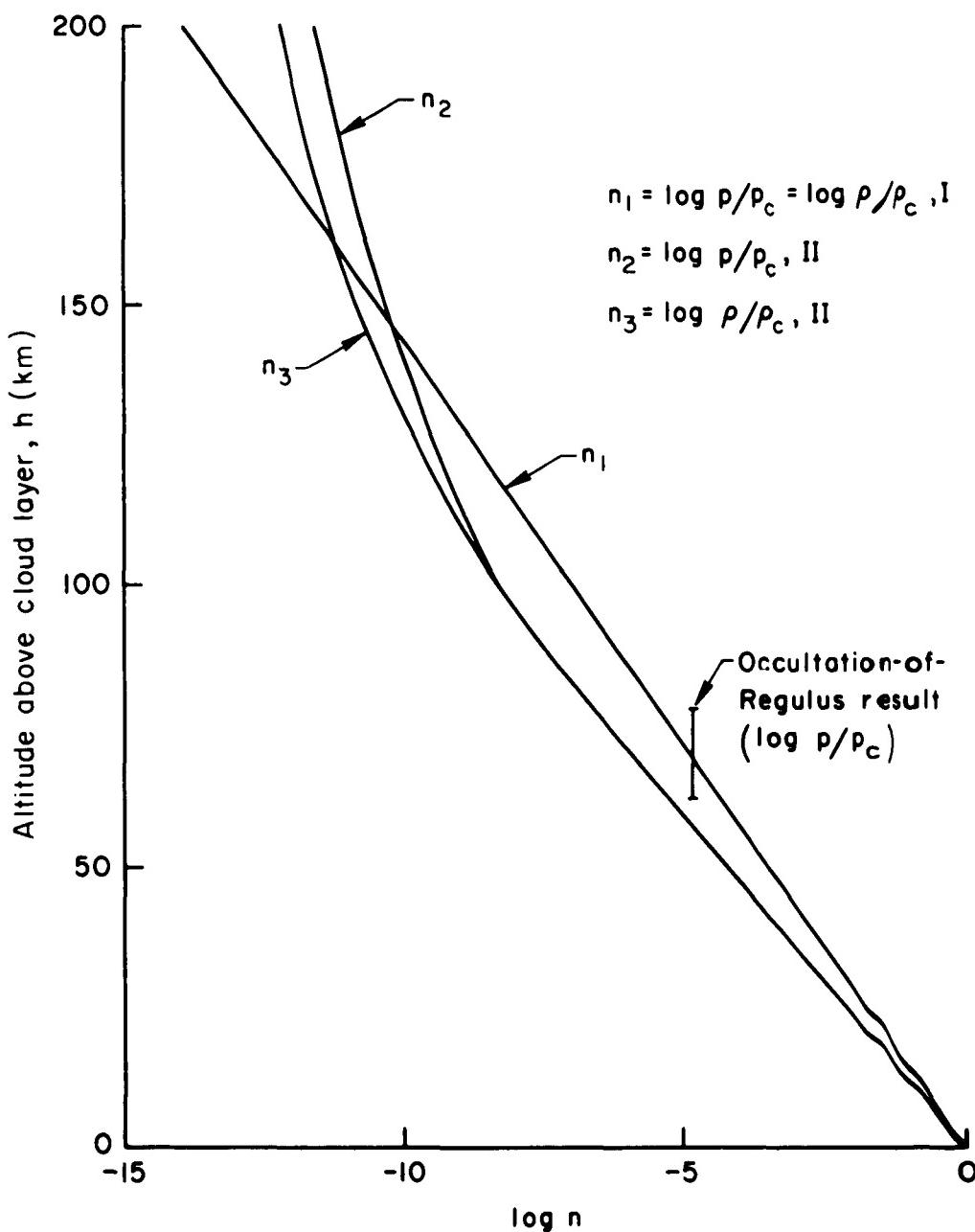
Fig. 4—Upper-atmosphere conditions for
CO_2 Models I and II ($h_c = 0$)
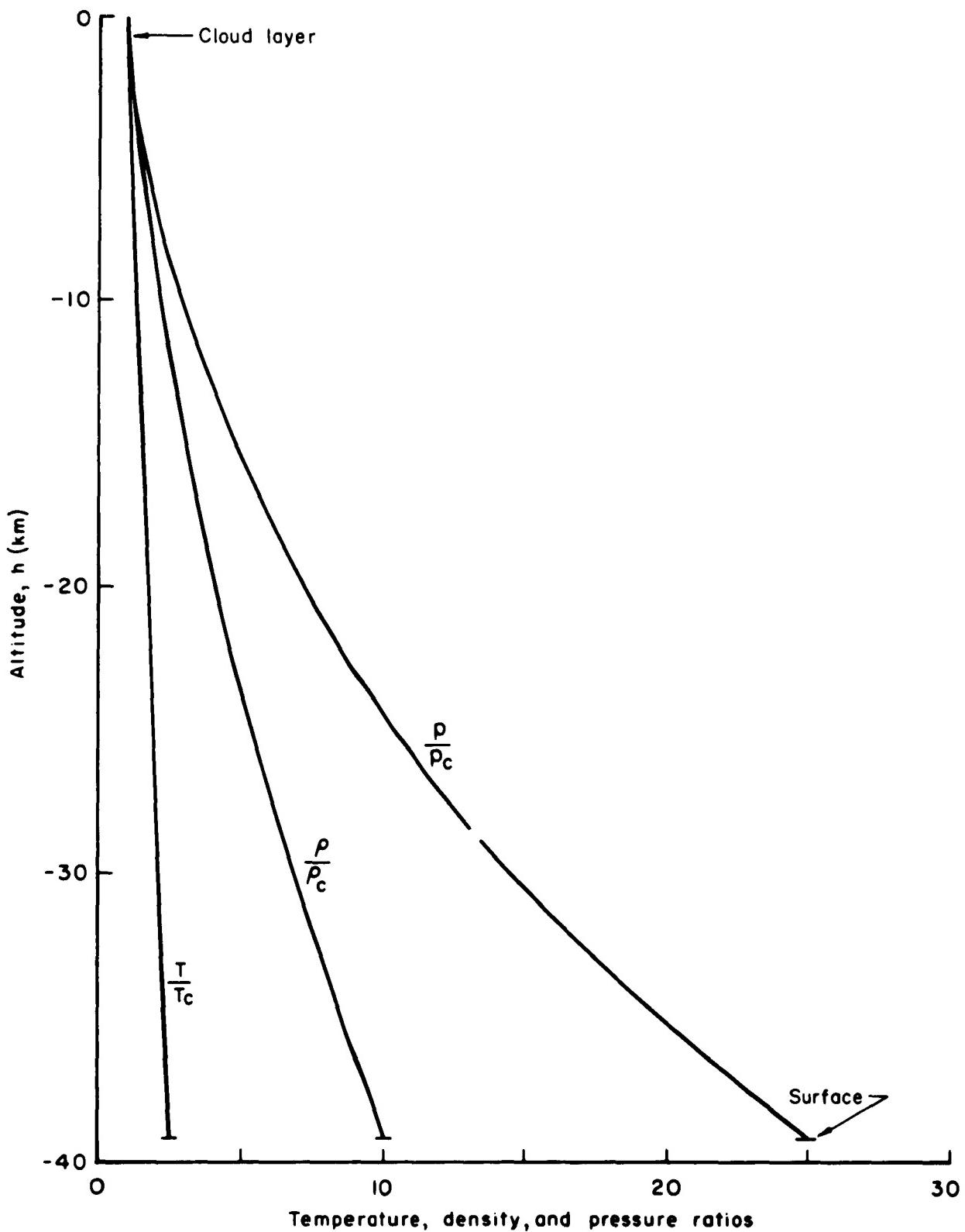
Fig. 5 — Lower atmospheric structure of
N_2 Model .I (T = 580 °K)

is based on the lower surface temperature of $350^{\circ}K$ (Figs. 1 and 6).
The tropopause temperature for both variants is $235^{\circ}K$. Again the expect-
ed differences in the troposphere are great. Calculations with 80 per
cent N_2 and 20 per cent CO_2 show negligible differences. The tropo-
sphere of N_2 Model II for Venus is most similar to that of the Earth;
the Cytherean model has a higher lapse rate (see Appendix A), and the
tropopause is somewhat higher above the surface.

The atmosphere above the tropopause is taken to be identical for
both N_2 models. The stratosphere shows a linear decrease in tempera-
ture to $190^{\circ}K$ at 60 km above the cloud level.[*] The heating rate in the
upper atmosphere is chosen arbitrarily to be less steep than that of
the CO_2 Model II, owing to the fact that since N_2 requires more energy
to be dissociated, less dissociation is present. Again, ionization
is neglected. An approximate mean temperature of the upper atmosphere
is about $285^{\circ}K$. The resulting structure above the cloud level is shown
in Fig. 7; the corresponding mean free path is shown in Fig. 8. The
curve showing mean free path as a function of altitude is close to that
of Earth. The Regulus result is met in the upper atmosphere.[**] Dis-
sociation effects are neglected as far as composition changes are con-
cerned.

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[*]Private communication from Lewis D. Kaplan.

[**]These values, however, have not been recomputed according to
the assumed compositon of the N_2 model.
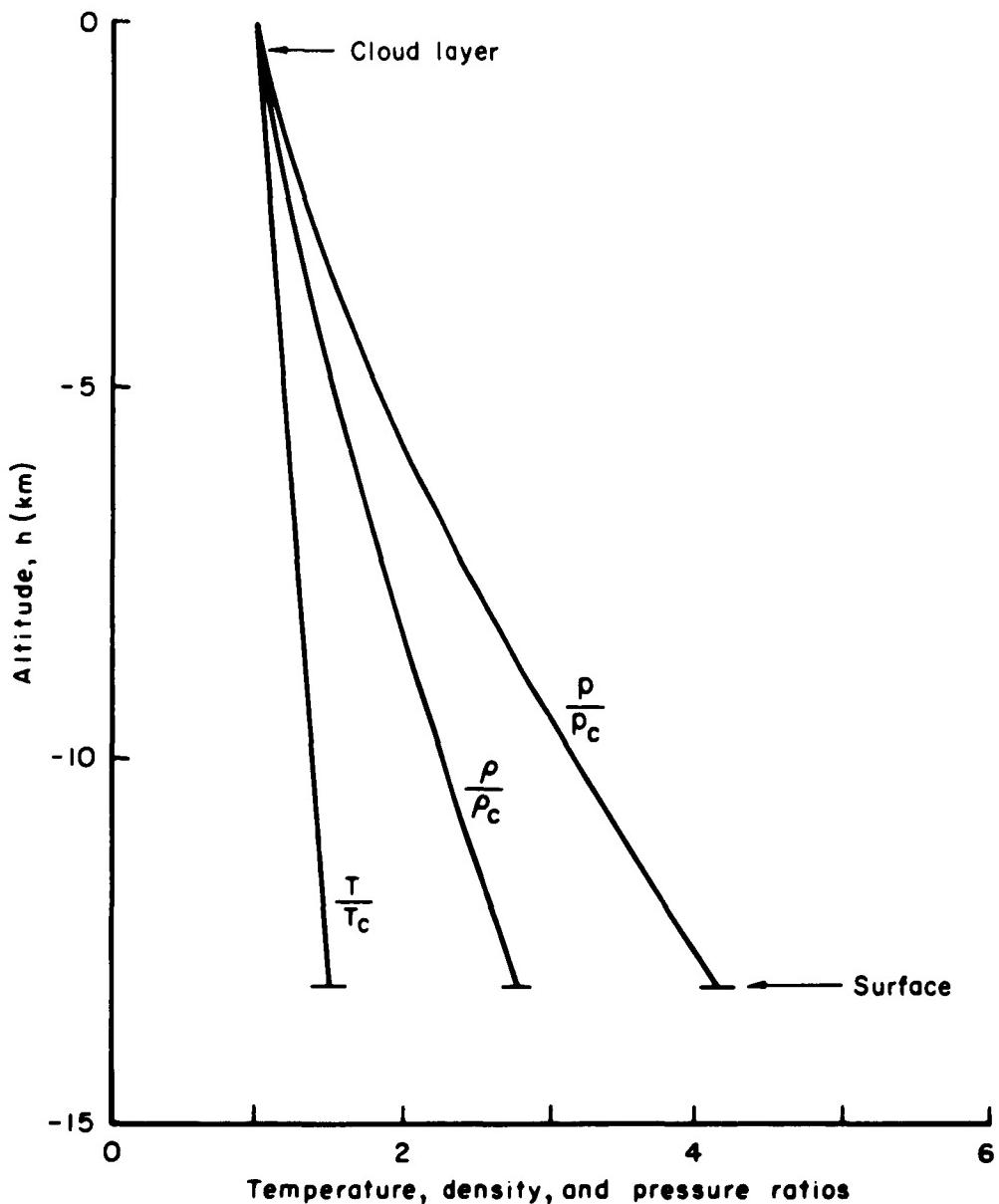
Fig. 6 — Lower atmospheric structure of
N_2 Model II $(T = 350 °K)$

$$n_1 = \log p / P_c$$
$$n_2 = \log P / P_c$$
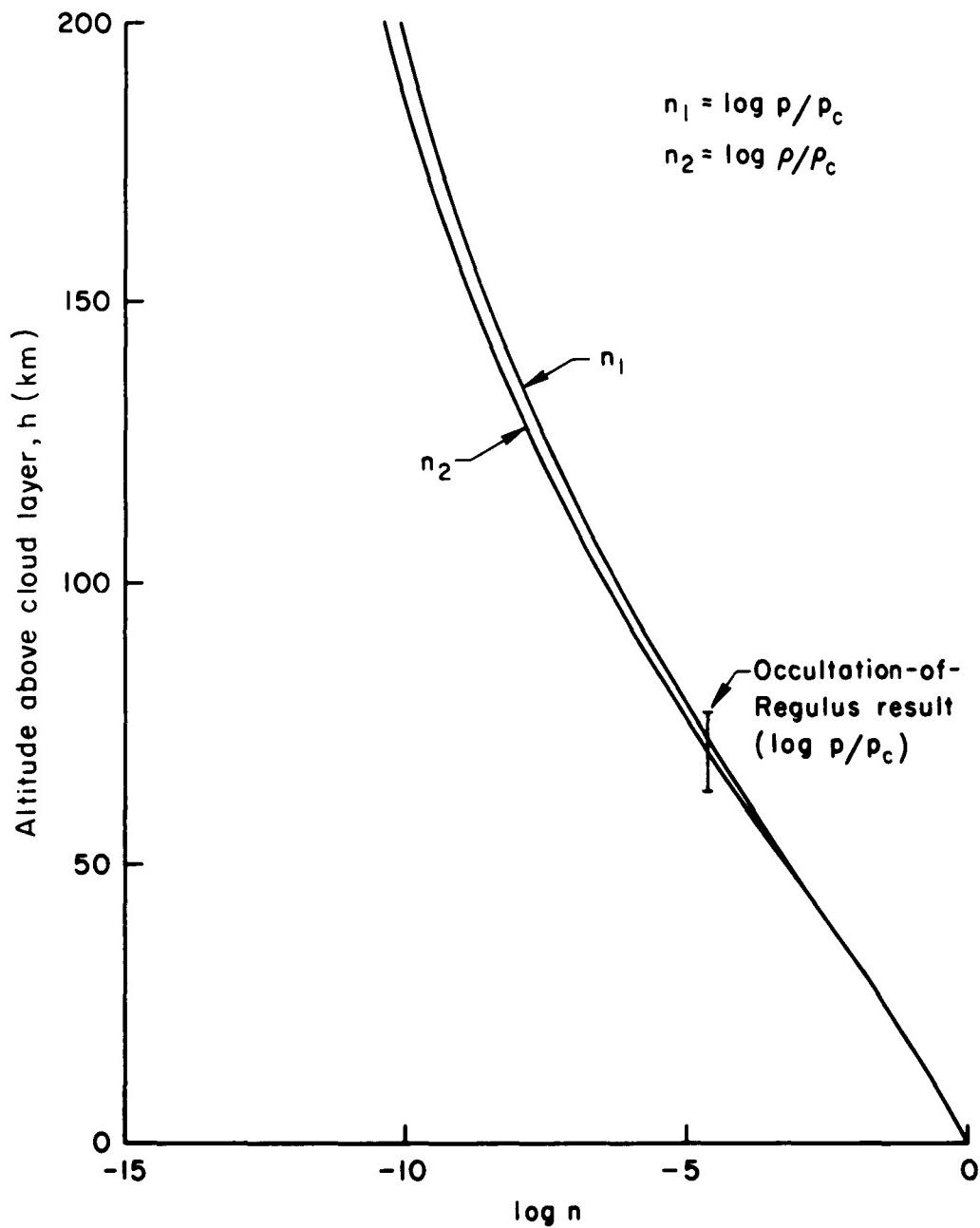
Fig. 7—Upper-atmosphere conditions for
N_2 Models I and II ($h_c = 0$)
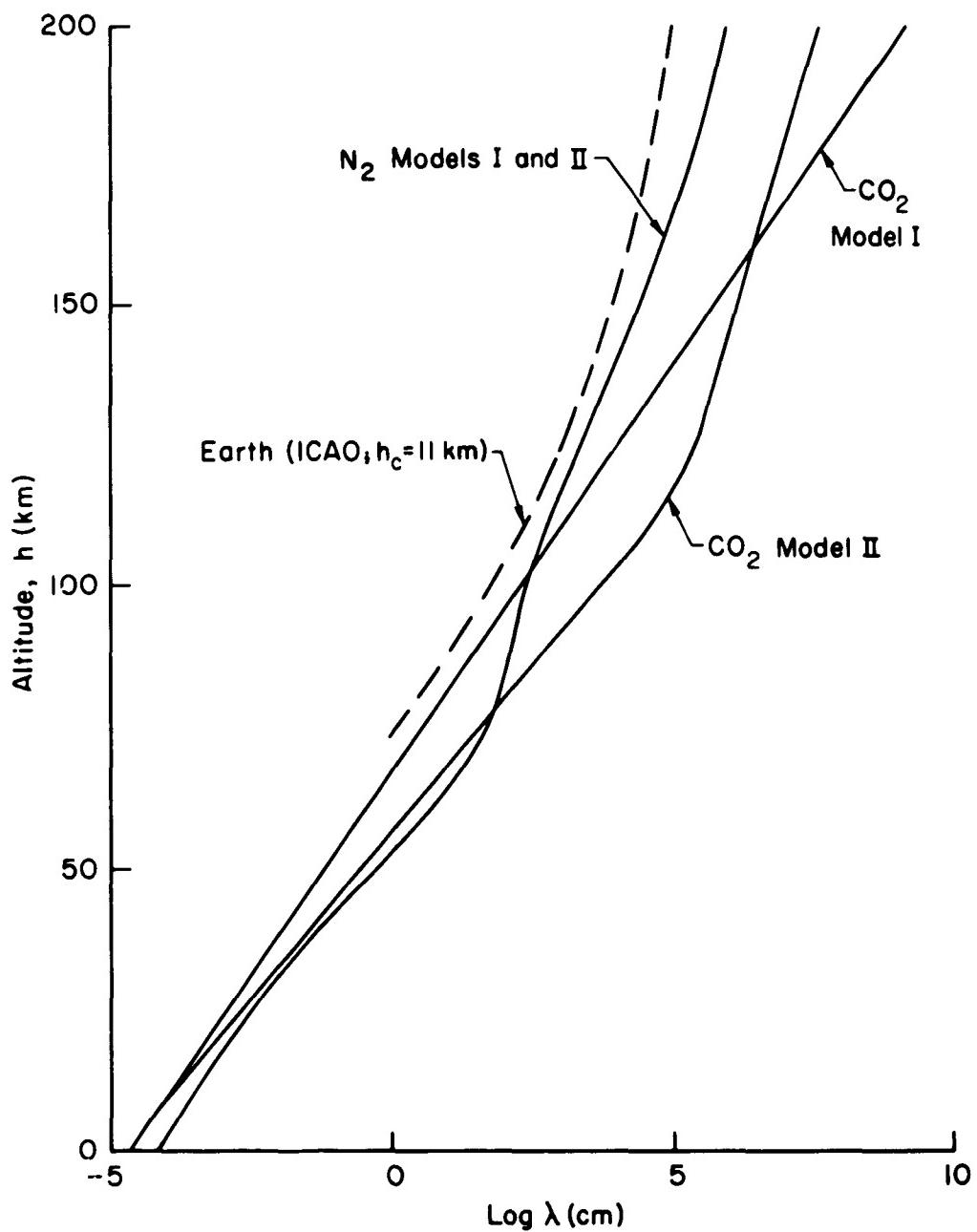
Fig. 8—Mean free path of all models and of Earth

IV. FLIGHT REGIMES

The flight regimes for all models of the Cytherean atmosphere
will be discussed in terms of free-stream Mach, Reynolds, and Knudsen
numbers:

$$M_\infty = \frac{\mu}{a_\infty}$$

$$Re_\infty = \frac{\mu \rho_\infty \ell}{\eta_\infty}$$

$$K_\infty = \frac{\lambda_\infty}{\ell}$$

A length ℓ of 1 m was chosen for all calculations; this length could
be the characteristic vehicle dimension or the nose radius of a blunt
body, for example, and all values given may readily be converted to a
different characteristic dimension. The speed of sound was taken to
be that for low frequencies, i.e., the acoustic velocity computed with
the vibrational mode of the molecules (if applicable) is in thermody-
namic equilibrium.[26] For blunt-body flows, the Reynolds numbers for
the free stream may be converted to those of the shock layer by keeping
in mind that the mass flow is constant and by considering the viscosity
change to be based only on the stagnation temperature determined by in-
cluding chemical reaction. However, finding local Mach numbers near
the body in reactive flow presents a difficult problem. The propaga-
tion speed of weak disturbances in chemically reactive media must be
considered in detail in order to select a correct value for the speed
of sound.[28] All viscosity values in this Memorandum were taken from
Ref. 26, and pressure effects on viscosity are neglected. For the N_2
model, viscosity values for pure N_2 were chosen as representative of
the gas mixture. In view of the other uncertainties it is felt that
a calculation of viscosity for the gas mixture is not warranted. The
Knudsen number, based on free-stream or incident-gas-mixture mean free
path in the M-Re plots, was computed from the relation

$$K_\infty = 2 \frac{a_\infty}{c_\infty} \frac{M_\infty}{Re_\infty} = \left(\frac{4}{3} \gamma \right)^{1/2} \frac{M_\infty}{Re_\infty}$$

A constant mean value of the ratio of the specific heats[26] was taken.
The mean free path was also found separately with the above equation
from

$$\lambda_\infty = \frac{\eta_\infty}{\rho_\infty a_\infty} \left(\frac{4}{3} \gamma \right)^{1/2}$$

The resulting values of these parameters for all models as a function
of altitude are given in Tables 1 to 3 in Appendix C.

Flight conditions in the troposphere are depicted in Figs. 9 and
10. For the models in which the high surface temperature is assumed,
high Reynolds numbers are found for given Mach numbers. This fact, in
conjunction with the fact that high ambient temperatures lead to high
stagnation temperatures, makes high-speed flight difficult indeed.
Conditions will be less severe in the troposphere for the N_2 model.
Also shown in Figs. 9 and 10 are the flight conditions for an accepted
surface temperature of $350^\circ K$. Flight Reynolds numbers for the CO_2
model are now in the range experienced on Earth; the N_2 model exhibits
smaller values. It is this drastic difference of flight states in the
troposphere, depending on surface temperatures accepted, which suggests
that the author should hold in abeyance a final decision on surface
temperature.

Figures 11 to 13 are traditional Mach number - Reynolds number
diagrams for all models above the cloud level. The scales on the figures
are identical, and the curves labeled for different altitudes repre-
sent the flight conditions of a 1-m object at the given altitudes. By
coincidence, flight conditions from the free-stream viewpoint are near-
ly identical for all models at about 80 km above the cloud layer (they
will of course be different in the shock layer because of the different
composition). The differences below this altitude are minor, since the
major departures occur at high altitude. This is to be expected in
view of the temperature distributions assumed in this range, and it
is not too important aerodynamically. The regime of transition from
free-molecule to continuum flow is indicated by the curves for free-stream
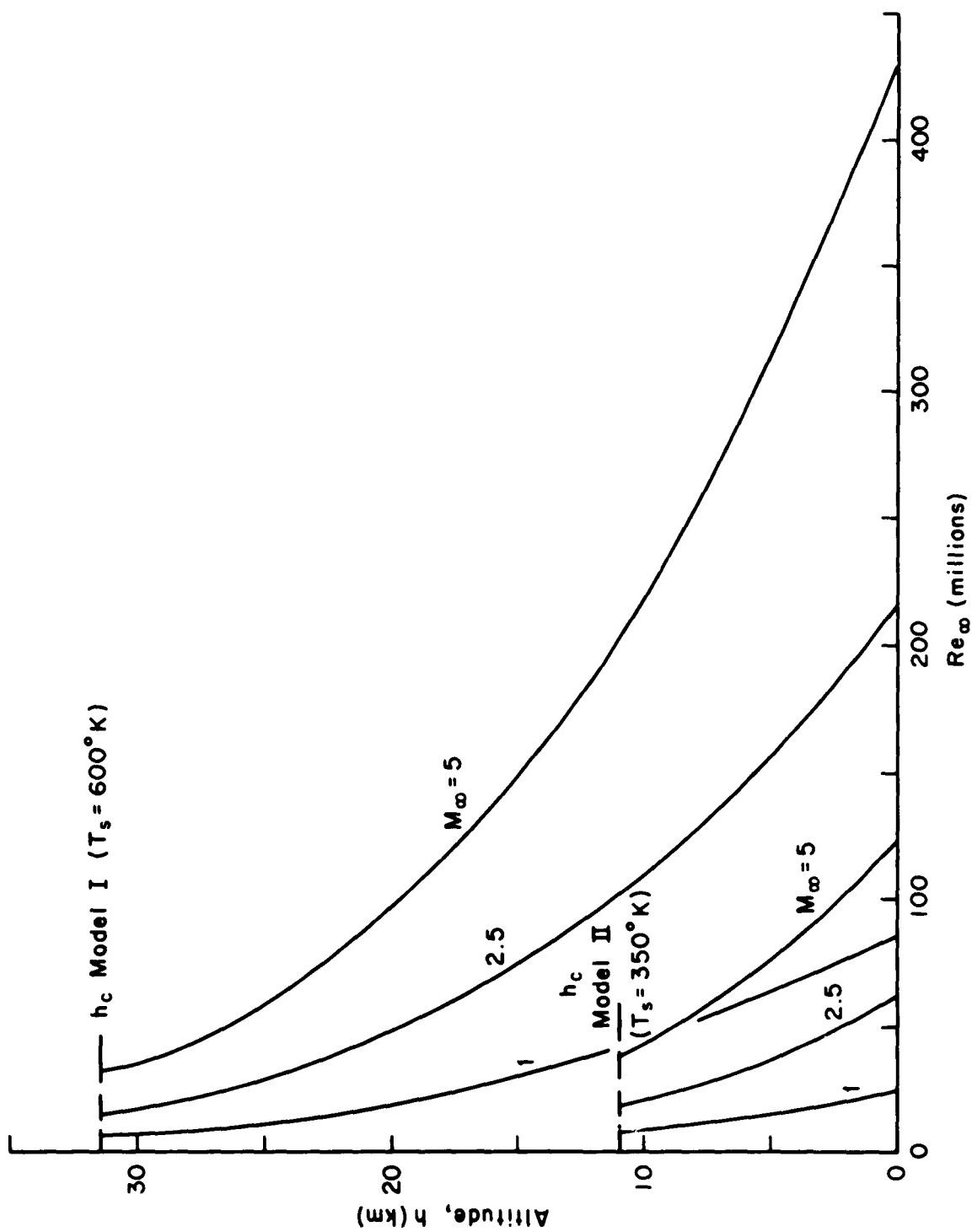
Fig. 9—Flight conditions in the troposphere of the CO_2 models ($\ell = 1$ m)
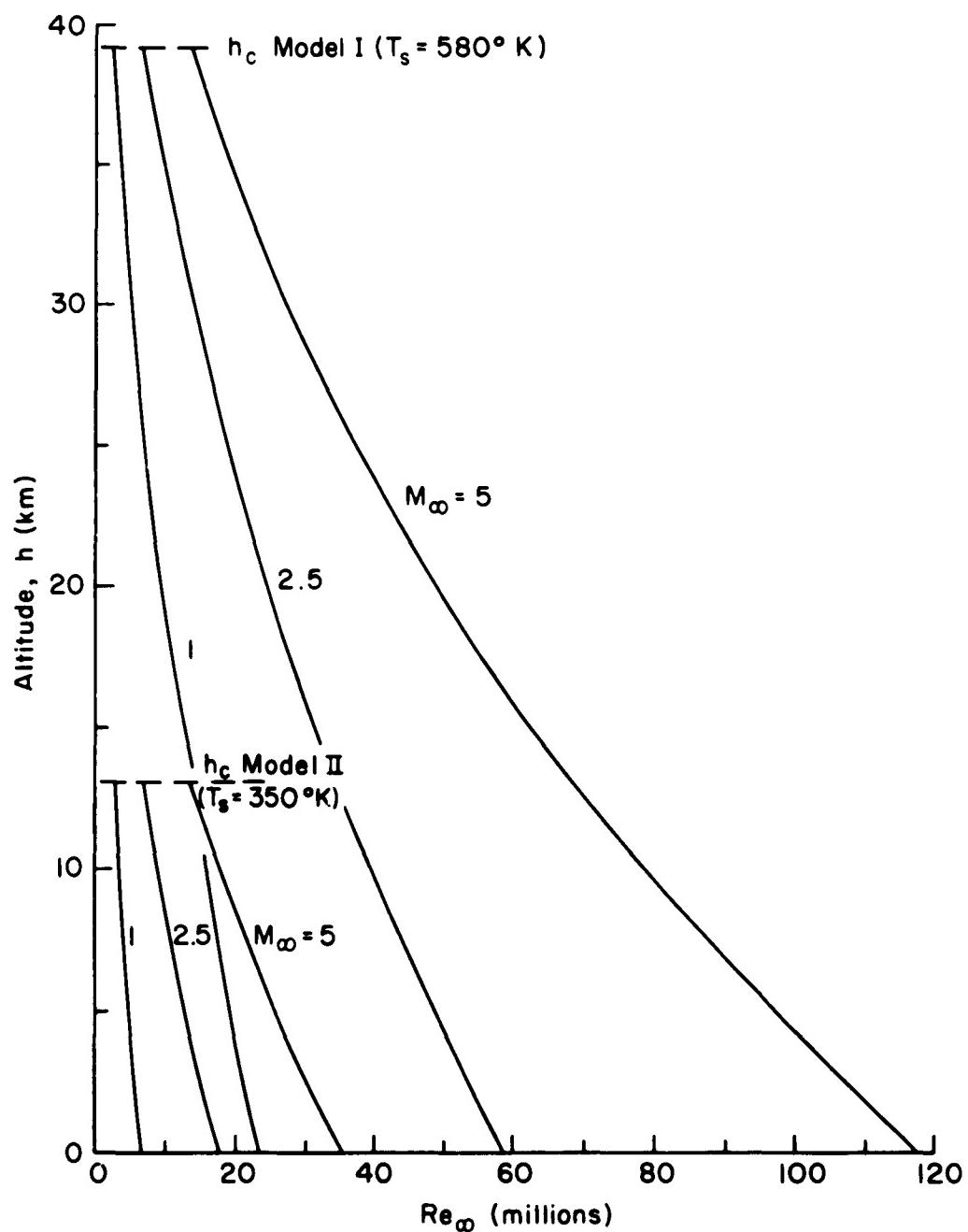
Fig. IO—Flight conditions in the troposphere of the
N_2 models (ℓ = Im)
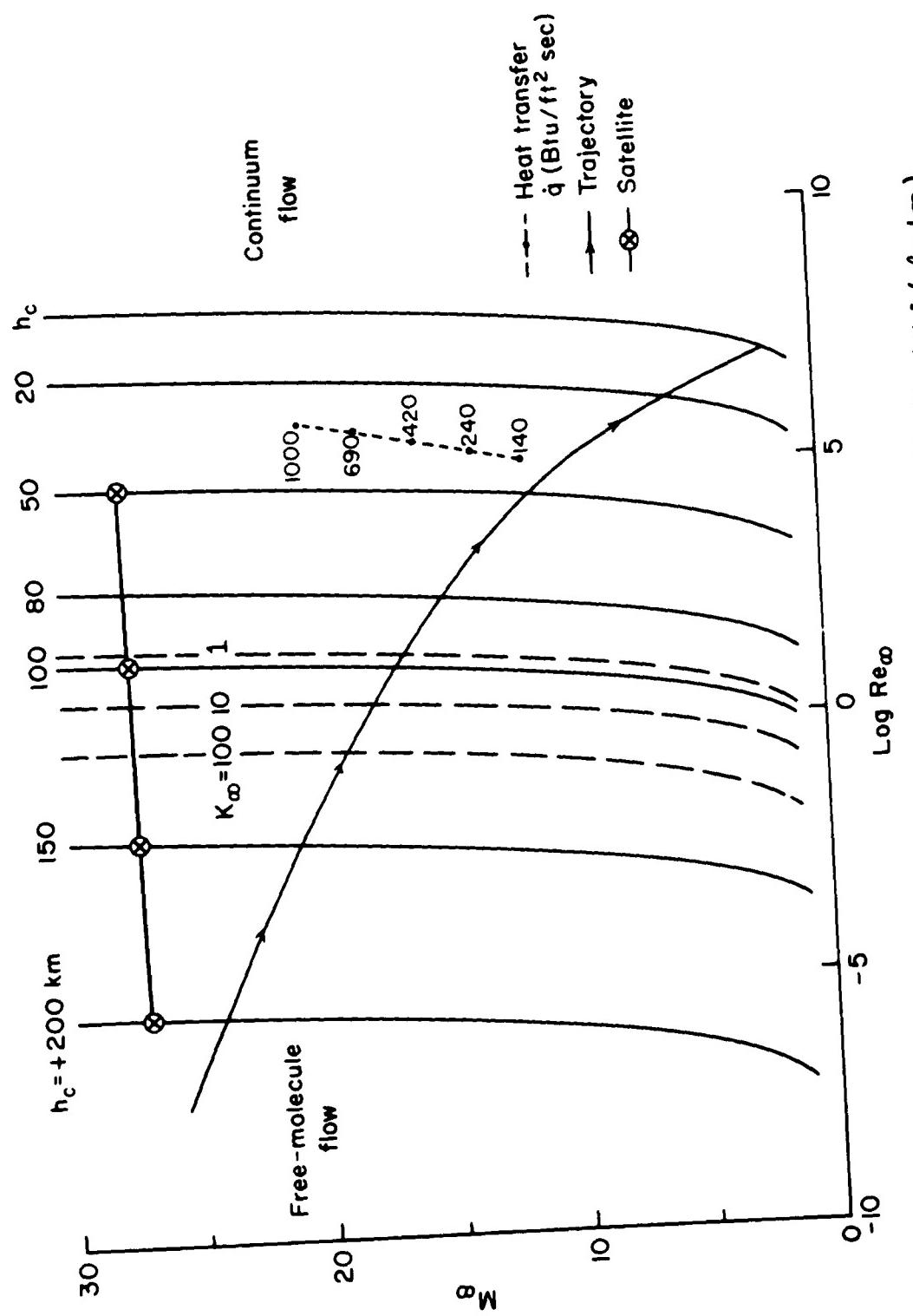
Fig. 11—Mach number versus Reynolds number for CO_2 Model I ($\ell = 1m$)
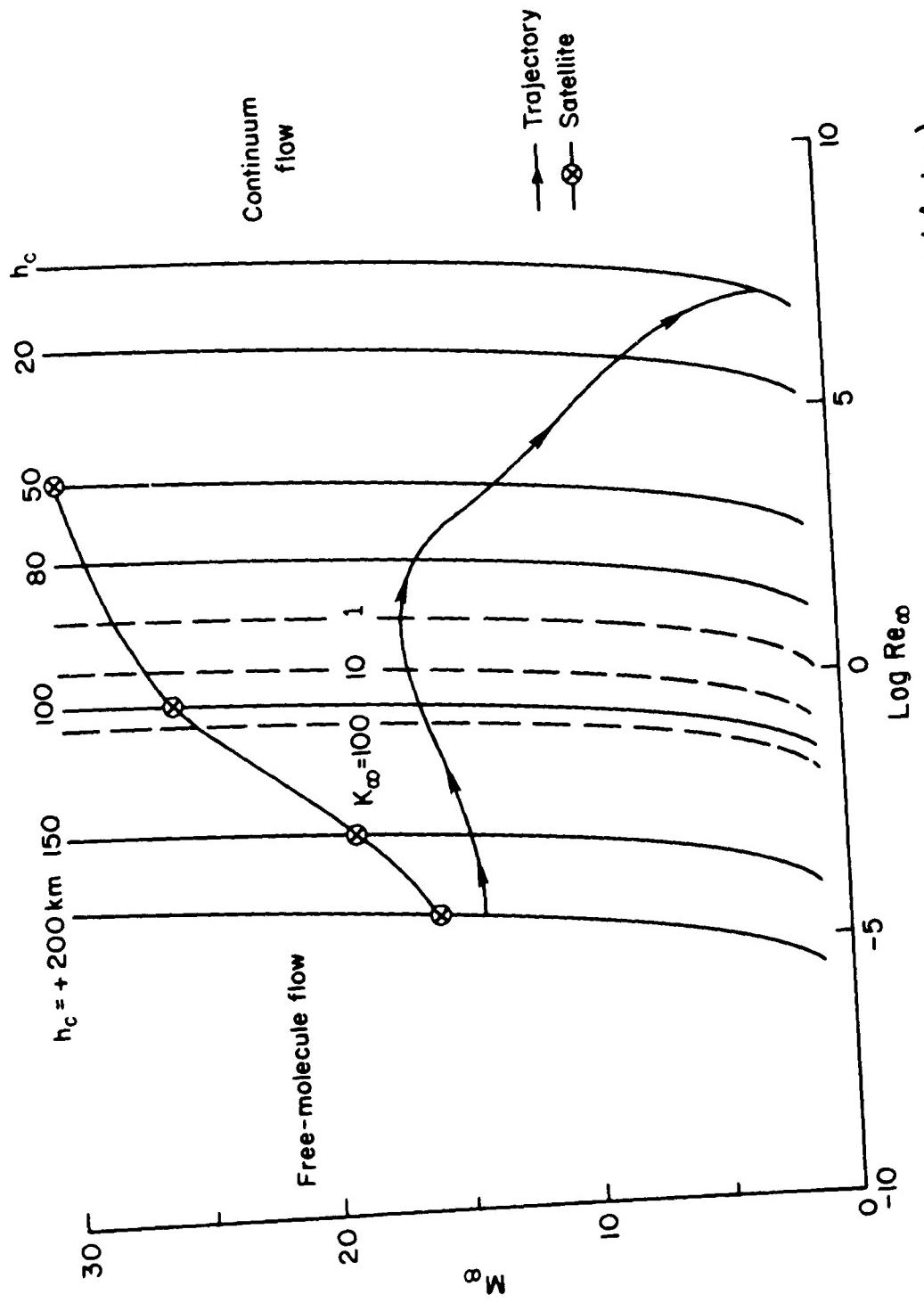
Fig. 12—Mach number versus Reynolds number for CO_2 Model II ($\ell = 1\,m$)
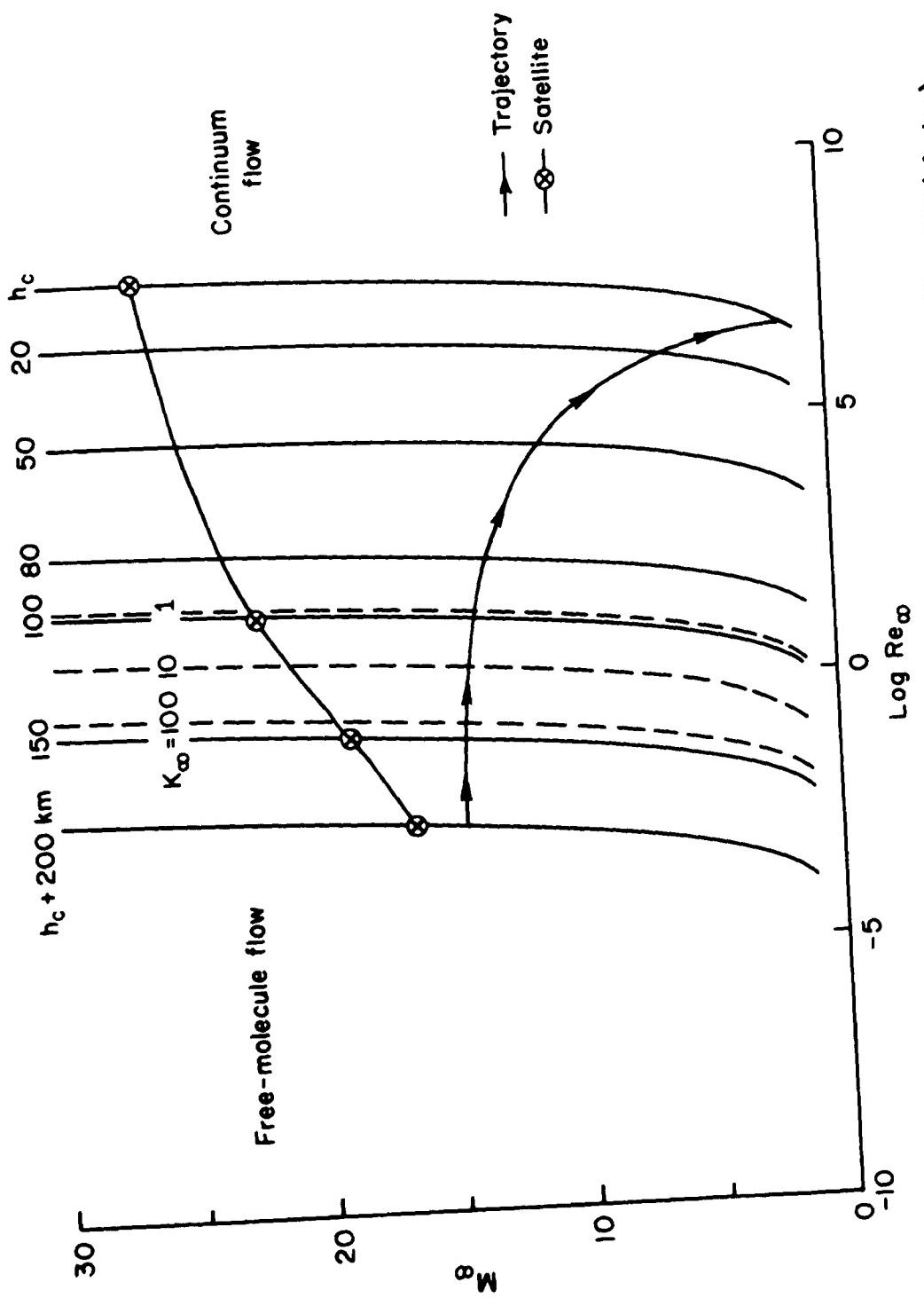
Fig. 13—Mach number versus Reynolds number for N_2 Models I and II ($\ell = 1m$)

Knudsen numbers. It is realized that it is "irrelevant"[25] simply
to choose a certain value of free-stream Knudsen number as a single
criterion for entering the transitional flow regime. The appropriate
mean free path to be selected depends on the problem to be investigated.
Generally speaking, the smallest typical mean free path in a given flow
situation serves as a criterion. However, there is no question that
the appearance of transitional-flow and continuum-flow phenomena are
loosely bounded by our given values of Knudsen number. If we assume
a hyperthermal flow*[29] in conjunction with the flow regime appli-
cable to the stagnation region of a blunt body, the following gener-
alizations are possible. Defining a detailed picture for the transitional
regime in the Earth's atmosphere for stagnation-point flow, Probstein
shows at which altitude and speed certain regimes are encountered.**
Comparing these regimes with the free-stream Knudsen number based on
the ICAO Standard Atmosphere for Earth, we find the approximate re-
lation of K_∞ to the type of flow field encountered. For $K_\infty = 100$, we
definitely deal with free-molecule flow. For $K_\infty = 10$, we enter the
so-called first-collision regime, or near-free-molecule flow; while
around $K_\infty = 1$, shock waves begin to form. These general findings are
likely to be valid generally for a planetary atmosphere, and they satis-
fy the current purpose of a general orientation. This picture is pre-
sented again ' Fig. 14. We see the following range of altitudes of
transition from free-molecule to continuum flow:

	Altitude (km)
CO_2 Model II	85 - 105***
CO_2 Model I	95 - 125***
N_2 Models I and II	95 - 145***
Earth	105 - 155****

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*In a hyperthermal flow, the flow speed is much larger than the mean
speed of the molecules in the free stream, and furthermore the body sur-
face with which the incoming molecules interact is highly cooled.

**See Fig. 1 of Ref. 25.

***Above cloud level.

****Above tropopause.
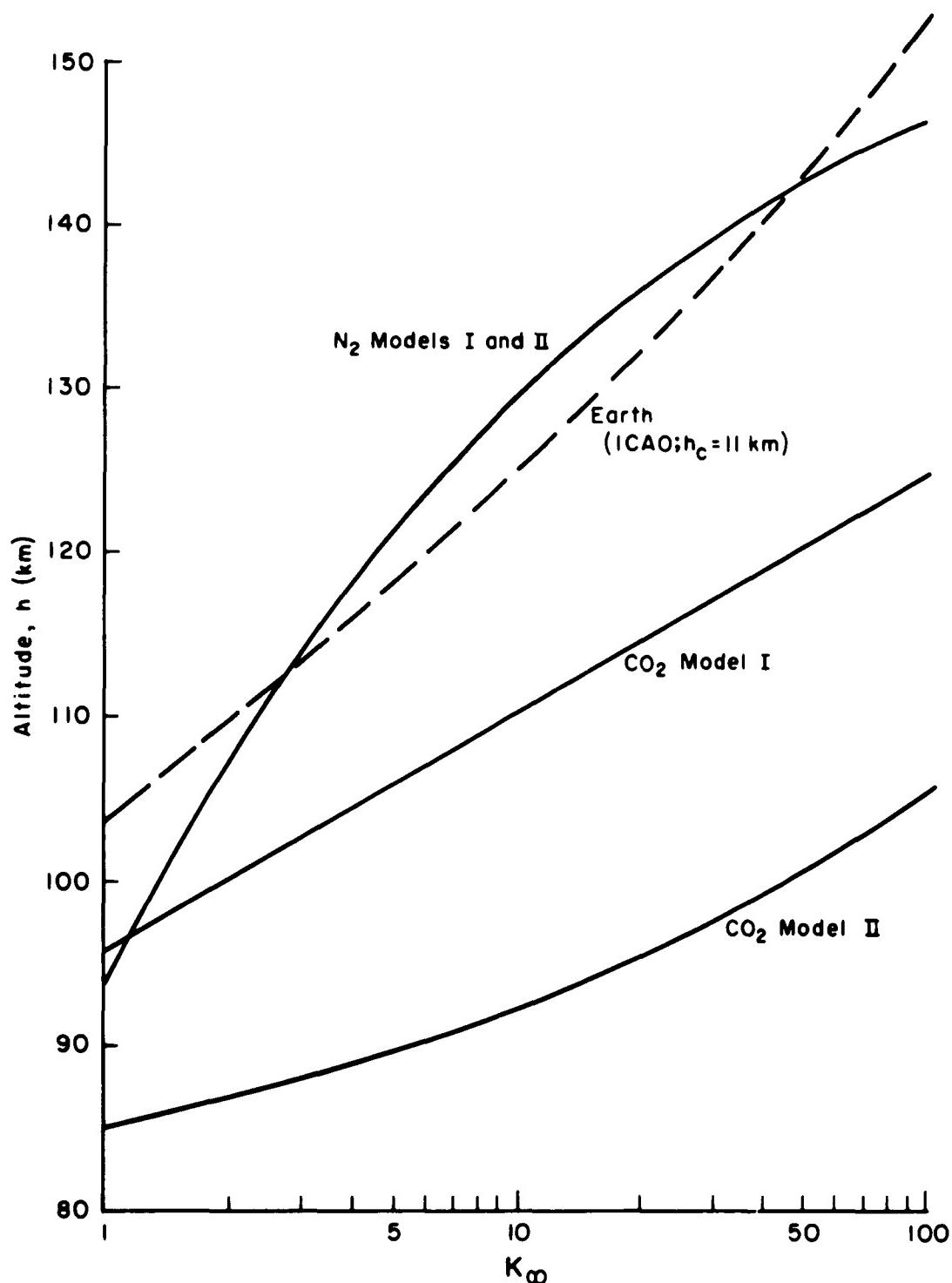
Fig. 14 — Knudsen number as a function of altitude
($h_c = 0$, $l = 1m$)

The CO_2 Model II represents the lowest altitude of transition; the N_2 model is closest to the situation on Earth, as expected. Again the approximate nature of these estimates needs to be emphasized.

V. AERODYNAMIC EXAMPLES

Some simple aerodynamic examples were chosen to illuminate further the difference in flight problems for the limiting atmospheric models. Satellite speeds were computed for different altitudes from

$$v = \mu = \left(rg_c \, \frac{r_c^2}{r^2} \right)^{1/2} , \qquad i.e., \quad \frac{v^2}{r} = g_c \left(\frac{r_c}{r} \right)^2$$

with $r = (h - h_c) + r_c$, and with r_c as the estimated radius of Venus at the top of the visible cloud layer. The resulting orbital speed is about 7.15 km/sec; it is nearly independent of altitude. Satellite flight conditions are indicated in Figs. 11 to 13. The Mach numbers differ widely at high altitudes owing to the variation in the speed of sound for the temperatures assumed (Fig. 1).

Furthermore, the flight path for a simple trajectory has been computed. It was assumed that an entry body with a characteristic dimension of 1 m enters the atmosphere vertically. The body is uniformly decelerated by unspecified means to attain a speed of 0.526 km/sec at the cloud level ($M_\infty \simeq 2$). The deceleration is 0.1 km/sec^2 or approximately 10 g on Earth. The total time of flight from h_c + 200 km to h_c, i.e., for 200 km, is about 1 min. The resulting flight conditions are also shown in Figs. 11 to 13. Free-stream Mach and Reynolds numbers as a function of altitude are given separately in Fig. 15. The differences in Mach number exhibited by the limiting models of the Cytherean atmosphere above about h_c + 100-km altitude are again striking. Certainly the exponential CO_2 atmosphere must be considered to be the less-realistic one. The transitional regime can be inferred from Figs. 14 and 15. It is seen that in the continuum flow, with its high aerodynamic forces, a relatively small range of conditions is encompassed for all models.

Shown in Fig. 11 are individual values of estimated heat-transfer rate to the stagnation point of a sphere of 1-m nose radius found in pure CO_2. These values were recomputed from the experimental results given by Yee et al.[30] obtained in a ballistic range with small models (nose radius = 0.11 in.) in pure CO_2. The conversion of the results
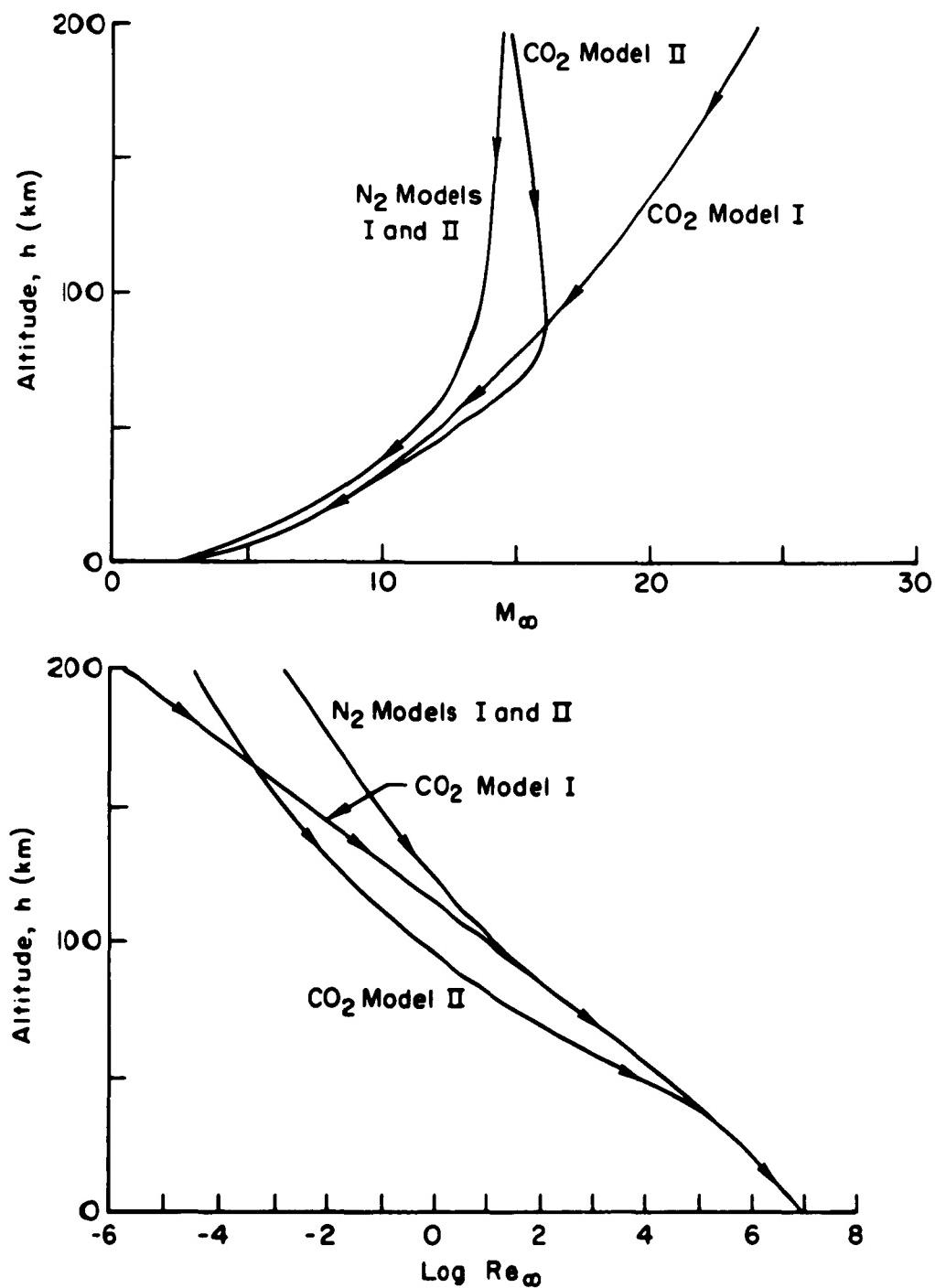
Fig. 15 — Trajectories of 1-m entry body for all models

to the much larger 1-m nose radius based on stagnation-point heat-transfer theory may not be applicable. At any rate, these values do include all the chemical effects expected in CO_2.

VI. CONCLUSIONS AND OUTLOOK

It is found that the best evidence available concerning the atmosphere of Venus does not permit an estimate of a single likely model atmosphere suitable for aerodynamic use. Taking the possible variation of the astronomer's results into account, it is seen that substantial differences in flight regimes must be considered. The largest discrepancies apply to the troposphere. A resolution of this unsatisfactory state of affairs requires a definitive answer on two major problems:

1. Surface temperature of Venus

2. Quantitative composition of the Cytherean atmosphere

From the free-stream viewpoint of flight, these answers will provide a single working basis for aerodynamic calculations. As is shown in this Memorandum, it is currently required that major variations be taken into account. However, it is believed to be unlikely that the actual Cytherean atmosphere will be outside the limits given. In each instance it is seen that the continuum flight regime is entered at greater altitude than 100 km or so above the visible cloud layer.

The author believes that a more detailed analysis of flow problems in shock layers, boundary layers, etc., would be premature at present. Knowledge of quantitative composition, including minor constituents, is required for such calculations. A complete analysis of reactive flows, including relaxation phenomena, has not yet been presented for the Earth's atmosphere; in view of the uncertainties pertaining to Venus, such effort is not now warranted for the Cytherean atmosphere(s). It is felt, however, that the limits given may provide design criteria for probes approaching the planet or possibly those reaching the cloud layer itself.

Appendix A

METHODS OF CALCULATION OF ATMOSPHERIC STRUCTURE

Calculations of thermodynamic parameters in a stable atmosphere are based on the fundamental equation of hydrostatics:

$$dp = -g\rho dh \tag{1}$$

The acceleration due to gravity, g, is in general a continuous function of altitude, because of the spherical shape of the planets. Its value decreases with distance from the surface according to

$$g = g_c \frac{r_c^2}{(r_c + h)^2} \tag{2}$$

where r_c is the radius of the planet (for Venus, measured to the top of the visible cloud layer), and g_c the value of g at the same point. In the case of Venus, r_c is still in doubt, and in general $h \ll r_c$. Therefore setting $g = g_c = $ const will result in small errors, in view of the other uncertainties of the properties of planetary atmospheres.

Pressures are sufficiently low in planetary atmospheres that the gases or gas mixtures may be treated as thermally perfect gases with

$$\rho = \frac{p\mu}{RT} \tag{3}$$

as long as no dissociation occurs. The mean molecular weight μ at a given altitude may be computed from the assumed composition by

$$\mu = \sum_i x_i \mu_i \tag{4}$$

where x_i and μ_i are mole fraction and molecular weight, respectively, of the ith species.

UNIFORM ATMOSPHERE AND THE CONCEPT OF SCALE HEIGHT

Assuming the density of an atmosphere to be constant for $h_1 < h < h_2$, Eq. (1) may be integrated immediately to give

$$H_o \overset{d}{=} h_2 - h_1 = \frac{1}{g\rho} \int_{p_2}^{p_1} dp = \frac{1}{g\rho} (p_1 - p_2) \tag{5}$$

Ordinarily the atmosphere cannot be realistically described by Eq. (5), excepting calculations for small altitude intervals, $\Delta h = h_2 - h_1$. However, the concept of "scale height" denoted by H_o has been derived from Eq. (5). Setting $h_2 - h_1 = H_o$, $p_2 = 0$, $p_1 = p_o$, and $\rho = \rho_o = $ const, we find from Eqs. (3) and (5)

$$H_o = \frac{p_o}{g\rho_o} = \frac{RT_o}{\mu g} \tag{6}$$

The density at standard (or normal) pressure and temperature is chosen for ρ_o.[*] The scale height, H_o, is a quantity which may be derived from the reduction of results from absorption spectra, since H_o is, in effect, proportional to the number of molecules causing the absorption in the ray of radiation. No statement on pressure and temperature distribution is then required, and the estimates of amounts of certain gases present are given as H_o in terms of length, usually km or cm at STP. The pressure at the bottom of such a "standard column" of gas of height H_o is given from Eq. (6) by

$$p_o = H_o g \rho_o \tag{7}$$

It needs to be noted that this pressure then refers to the reflecting layer (or reversal layer) for the wavelength from which H_o has been

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[*]This state is sometimes abbreviated by STP or NTP. Pressure refers to 760 mm Hg; however, depending on authors, either $0^\circ C$ or $15^\circ C$ is chosen as standard temperature.

derived by absorption measurements. This reflecting layer does not necessarily coincide with the planet's surface or with a "cloud layer."

ISOTHERMAL ATMOSPHERE

For the isothermal atmosphere, also called the exponential atmosphere, it is assumed that the temperature is constant with altitude-- $dT/dh = 0$. From Eqs. (1) and (3) we obtain

$$\frac{dp}{p} = - \frac{g\mu}{RT} \, dh \tag{8}$$

For $T = T_1$ = const, Eq. (8) may be integrated to give

$$\ell n \; p = - \frac{g\mu}{RT_1} \, h \; + \; const \tag{9}$$

If we assume $p = p_1$, at $h = h_1$ the integrational constant in Eq. (9) may be evaluated, and the result is the well-known barometric height formula

$$p = p_1 \; \ell xp \left[- \frac{g\mu}{RT_1} \, (h - h_1) \right] \tag{10}$$

from which altitude may be found from a pressure measurement. Since T_1 = const, $p/p_1 = \rho/\rho_1$.

For numerical calculation, Eq. (10) may be rewritten

$$\log_{10} \frac{p}{p_1} = - \; 0.434 \; \frac{g\mu}{RT_1} \, (h - h_1) \tag{11}$$

ADIABATIC AND POLYTROPIC ATMOSPHERE

In this case the properties are assumed to vary with

$$pv^\gamma = \frac{p}{\rho^\gamma} = const \tag{12}$$

and therefore temperature and pressure are related by Poisson's law

$$\frac{T}{T_1} = \left(\frac{p}{p_1} \right)^{\frac{\gamma-1}{\gamma}} \tag{13}$$

For a given interval of height, in order to avoid numerical integration of Eq. (8) with Eq. (13), the ratio of the specific heats $\gamma = c_p/c_v$ may be taken as constant. Again this is ordinarily permissible for estimates of the structure of planetary atmospheres because of the other uncertainties. Also, the assumption of constant composition, i.e., constant mean molecular weight, may be retained, excepting calculations at high altitudes. Finally, the use of Eq. (13) excludes condensation processes with local enthalpy changes, and the property change is therefore called "dry adiabatic."

From Eq. (8) we obtain

$$h - h_1 = \frac{R}{\mu g} \int_{p_2}^{p_1} \frac{T}{p} \, dp \tag{14}$$

where the integral may be solved, considering Eq. (13). Again evaluating the integrational constant for $p = p_1$, and $T = T_1$ (with T now being a variable), we find

$$h - h_1 = \frac{RT_1}{g\mu} \frac{\gamma}{\gamma - 1} \left[1 - \left(\frac{p}{p_1} \right)^{\frac{\gamma-1}{\gamma}} \right] \tag{15}$$

Solving Eq. (15) for pressure, with $h_1 = 0$, gives

$$p = p_1 \left(1 - \frac{\gamma - 1}{\gamma} h \frac{g\mu}{RT_1} \right)^{\frac{\gamma}{\gamma-1}} \tag{16}$$

suitable for direct calculation. Equation (16) may be rewritten by defining the "dry adiabatic lapse rate"

$$\beta \equiv \left(\frac{\partial T}{\partial h} \right)_{ad} = - \frac{\gamma - 1}{\gamma} \frac{g\mu}{R} = - \frac{g}{c_p} \left(\frac{\text{temperature}}{\text{length}} \right) \tag{17}$$

to give

$$p = p_1 \left(1 + \beta \frac{h}{T_1}\right)^{\frac{\gamma}{\gamma-1}} \tag{18}$$

It is seen that in a dry adiabatic atmosphere, temperature decreases linearly with altitude, since the lapse rate is a constant. Inserting Poisson's law (Eq. (13)) in Eq. (15) to find the temperature, we have

$$T = T_1 + \beta h \tag{19}$$

or for altitude

$$h = \frac{T - T_1}{\beta} \tag{20}$$

Finally we may derive an expression for the ratio of the specific heats from the definition of the lapse rate (Eq. (17)) by

$$\gamma = - \frac{1}{1 + \beta \frac{R}{g\mu}} \tag{21}$$

Equations (15), (16), (18), (19), (20), and (21) may now be used in varying ways to estimate the structure of the adiabatic portion of planetary atmospheres. For example, one may assume a certain reasonable dry adiabatic lapse rate first and find γ from Eq. (21). This γ must now be compatible with the thermodynamic properties for the given gas or gas mixture in the range of pressures and temperatures under discussion. Or, in turn, the lapse rate to be expected may be computed, based on thermodynamic properties and Eq. (17). If, for example, the temperature at the tropopause and the surface has been estimated, the height of the troposphere may be found from Eqs. (17) and (20) for any choice of the lapse rate. More combinations of known parameters may thus be fitted into a consistent atmospheric model. In order to obtain atmospheric stability at the boundaries of an adiabatic and iso-

thermal structure of an atmosphere, for example, temperature and pressure as well as dp/dh must be continuous. Similar expressions are given in the meteorological literature for moist adiabatic atmospheres, including the effects of water vapor in local thermodynamic equilibrium. The lapse rates are smaller in this instance than in the dry case, because of the heating of the atmosphere from the heat of condensation.

Finally, if from some observation the temperature is given as a function of altitude, and the pressure is also known independently at some point, the equations may be used formally by replacing γ by a polytropic exponent m. In this manner, increases of T with h may also be treated to obtain the function $p(h)$, etc. Such an atmosphere may then be called polytropic.

Appendix B

MODELS OF THE VENUS ATMOSPHERE

CO_2 MODEL I

General data:

Composition, etc.

100 per cent CO_2

$\mu = 44$, $g = 860$ cm/sec^2

$\frac{g\mu}{R} = 45.5$ °K/km, $\frac{R}{g\mu} = 0.0220$ km/°K

a, low frequency speed of sound as $f(T,p)$

η, viscosity of CO_2 at 1 atm

Surface to cloud layer: $h_s < h < h_c$

Dry adiabatic structure, $\gamma = 1.25 = $ const

$\beta = -10$°K/km

$h_c - h_s = 31.5$ km

$p_s = 7.02$ atm, $T_s = 600$°K

$p_c = 0.170$ atm, $T_c = 285$°K

Cloud layer to exosphere: $h_c < h < (h_c + 200$ km)

Isothermal structure, dissociation neglected

$\gamma = 1.29$ (for calculation of K, etc.)

$T = T_c = 285$°K $ = $ const

$\log \frac{p}{p_c} = \log \frac{\rho}{\rho_c} = -0.0693$ h for h in km

CO_2 MODEL II

General data:

Composition, etc.

100 per cent CO_2

$\mu = 44$, $g = 860$ cm/sec^2

$\frac{g\mu}{R} = 45.5$ $^\circ$K/km, $\frac{R}{g\mu} = 0.220$ km/$^\circ$K

a, low frequency speed of sound as $f(T,p)$

η, viscosity of CO_2 at 1 atm

Surface to cloud layer: $h_s < h < h_c$

Dry adiabatic structure, $\gamma = 1.30 = $ const

$\beta = -10^\circ$K/km

$h_c - h_s = 11$ km

$P_s = 0.955$ atm, $T_s = 350^\circ$K

$P_c = 0.170$ atm, $T_c = 235^\circ$K

Above cloud layer: $h_c < h < (h_c + 85$ km$)$

Isothermal structure

$T = T_c = 235^\circ$K $= $ const

$\log \frac{p}{p_c} = \log \frac{\rho}{\rho_c} = -0.0840$ h for h in km

$$(h_c + 85 \text{ km}) < h < (h_c + 200 \text{ km})$$

Polytropic structure, dissociation neglected

$\beta = 6.14^\circ$K/km

$m = 0.881$

N_2 MODELS I AND II

General data I and II:

Composition, etc.

85 per cent N_2,   15 per cent CO_2

μ = 30.4,   g = 860 cm/sec^2

$\frac{g\mu}{R}$ = 31.4 $^\circ$K/km, $\frac{R}{g\mu}$ = 0.0318 km/$^\circ$K

$a = \left(\frac{\gamma R T}{\mu} \right)^{1/2}$

η, viscosity of N_2 at 1 atm

Surface to cloud layer: $h_s < h < h_c$

I. Dry adiabatic structure, γ = 1.39 = const

β = -8.81 $^\circ$K/km

$h_c - h_s$ = 39.2 km

p_s = 2.21 atm, T_s = 580°K

p_c = 0.0888 atm, T_c = 235 $^\circ$K

II. Dry adiabatic structure, γ = 1.39 = const

β = -8.81 $^\circ$K/km

$h_c - h_s$ = 13.1 km

p_s = 0.367 atm, T_s = 350°K

p_c = 0.0888 atm, T_c = 235°K

I and II above cloud layer: $h_c < h < (h_c + 60$ km$)$

Polytropic structure for

β = -0.75°K/km,   m = 1.024

$(h_c + 60$ km$) < h < (h_c + 200$ km$)$

Polytropic structure for

β = 2.21°K/km,   m = 0.934

Appendix C

THERMODYNAMIC PROPERTIES OF CO_2 AND N_2 MODELS

Table 1

CO_2 MODEL I

h (km)	T (°K)	p (atm)	ρ (g/cm³)	$\frac{p}{p_c}$	$\frac{\rho}{\rho_c}$	$\log \frac{p}{p_c}$	$\log \frac{\rho}{\rho_c}$	a (m/sec)	$\frac{g}{sec\ cm}\eta \times 10^{-4}$	λ (cm)	$\frac{Re}{M}$[a]
-31.5	600	7.02	6.28×10^{-3}	41.3	19.6	1.62	1.29	369	2.69	1.52×10^{-6}	8.61×10^{7}
-30	585	6.18	5.67×10^{-3}	36.4	17.7	1.56	1.27	365	2.63	1.63×10^{-6}	7.87×10^{7}
-25	535	3.95	3.96×10^{-3}	23.2	12.4	1.37	1.09	351	2.45	2.26×10^{-6}	5.67×10^{7}
-20	485	2.41	2.67×10^{-3}	14.2	8.34	1.15	0.921	335	2.26	3.25×10^{-6}	3.96×10^{7}
-15	435	1.40	1.73×10^{-3}	8.23	5.40	0.915	0.732	318	2.07	4.83×10^{-6}	2.66×10^{7}
-10	385	0.758	1.06×10^{-3}	4.46	3.31	0.649	0.520	301	1.88	7.61×10^{-6}	1.70×10^{7}
-5	335	0.380	6.10×10^{-4}	2.24	1.91	0.350	0.281	283	1.66	1.25×10^{-5}	9.04×10^{6}
0	285	0.170	3.20×10^{-4}	1	1	0	0	263	1.43	2.23×10^{-5}	5.88×10^{6}
20	285	6.99×10^{-3}	1.31×10^{-5}	4.11×10^{-2}	4.11×10^{-2}	-2.39	-1.39	263	1.43	5.44×10^{-4}	2.41×10^{5}
50	285	5.83×10^{-5}	1.10×10^{-7}	3.43×10^{-4}	3.43×10^{-4}	-3.47	-3.47	263	1.43	6.48×10^{-2}	2.02×10^{3}
80	285	4.86×10^{-7}	9.44×10^{-10}	2.86×10^{-6}	2.86×10^{-6}	-5.54	-5.54	263	1.43	7.55	1.74×10
100	285	2.01×10^{-8}	3.78×10^{-11}	1.18×10^{-7}	1.18×10^{-7}	-6.93	-6.93	263	1.43	1.89×10^{2}	6.95×10^{-1}
150	285	6.77×10^{-12}	1.27×10^{-14}	3.98×10^{-11}	3.98×10^{-11}	-10.4	-10.4	263	1.43	5.61×10^{5}	2.34×10^{-4}
200	285	2.35×10^{-15}	4.42×10^{-18}	1.38×10^{-14}	1.38×10^{-14}	-13.9	-13.9	263	1.43	1.61×10^{9}	8.13×10^{-8}

[a] For l = 1 m.

Table 2

CO_2 MODEL II

h (km)	T (°K)	p (atm)	ρ (g/cm³)	$\frac{p}{p_c}$	$\frac{\rho}{\rho_c}$	$\log \frac{p}{p_c}$	$\log \frac{\rho}{\rho_c}$	a (m/sec)	$\left(\frac{g}{\text{sec cm}} \times 10^{-4}\right)$	λ (cm)	$\frac{Re}{M}$[a]
-11	350	0.955	1.47×10^{-3}	5.62	3.79	0.750	0.579	288	1.72	5.28×10^{-6}	2.46×10^{7}
-8	319	0.638	1.07×10^{-3}	3.75	2.76	0.574	0.441	276	1.56	6.95×10^{-6}	1.87×10^{7}
-5	287	0.404	7.56×10^{-4}	2.38	1.95	0.376	0.290	263	1.44	9.45×10^{-6}	1.39×10^{7}
-2	256	0.247	5.18×10^{-4}	1.45	1.33	0.161	0.124	249	1.25	1.28×10^{-5}	1.03×10^{7}
0	235	0.170	3.88×10^{-4}	1	1	0	0	242	1.22	1.70×10^{-5}	7.70×10^{6}
20	235	3.55×10^{-3}	8.11×10^{-6}	2.09×10^{-2}	2.09×10^{-2}	-1.68	-1.68	242	1.22	8.14×10^{-4}	1.61×10^{5}
30	235	5.13×10^{-4}	1.17×10^{-6}	3.02×10^{-3}	3.02×10^{-3}	-2.52	-2.52	242	1.22	5.64×10^{-3}	2.32×10^{4}
40	235	7.43×10^{-5}	1.70×10^{-7}	4.37×10^{-4}	4.37×10^{-4}	-3.36	-3.36	242	1.22	3.88×10^{-2}	3.37×10^{3}
50	235	1.07×10^{-5}	2.45×10^{-8}	6.31×10^{-5}	6.31×10^{-5}	-4.20	-4.20	242	1.22	2.69×10^{-1}	4.86×10^{2}
60	235	1.55×10^{-6}	3.54×10^{-9}	9.12×10^{-6}	9.12×10^{-6}	-5.04	-5.04	242	1.22	1.86×10^{0}	7.02×10^{1}
70	235	2.24×10^{-7}	5.12×10^{-10}	1.32	1.32	-5.88	-5.88	242	1.22	1.29×10^{1}	1.02×10^{1}
80	235	3.25×10^{-8}	7.41×10^{-11}	1.91×10^{-7}	1.91×10^{-7}	-6.72	-6.72	242	1.22	8.91×10^{1}	1.47×10^{0}
85	235	1.23×10^{-8}	2.81×10^{-11}	7.25×10^{-8}	7.25×10^{-8}	-7.14	-7.14	242	1.22	2.35×10^{2}	5.58×10^{-1}
100	327	1.08×10^{-9}	1.77×10^{-12}	6.34×10^{-9}	4.56×10^{-9}	-8.20	-8.34	280	1.62	4.28×10^{3}	3.06×10^{-2}
120	450	1.03×10^{-10}	1.23×10^{-13}	6.04×10^{-10}	3.17×10^{-10}	-9.22	-9.50	323	2.13	7.02×10^{4}	1.87×10^{-3}
150	634	7.91×10^{-12}	6.70×10^{-15}	4.65×10^{-11}	1.73×10^{-11}	-10.3	-10.8	379	2.79	1.44×10^{6}	9.10×10^{-5}
180	818	1.21×10^{-12}	7.94×10^{-16}	7.12×10^{-12}	2.05×10^{-12}	-11.1	-11.7	427	3.37	1.30×10^{7}	1.01×10^{-5}
200	941	4.32×10^{-13}	2.47×10^{-16}	2.54×10^{-12}	6.37×10^{-13}	-11.6	-12.2	455	3.73	4.35×10^{7}	3.01×10^{-6}

[a]For $l = 1$ m.

Table 3A

N_2 MODELS I AND II

h (km)	T (°K)	p (atm)	ρ (g/cm³)	$\frac{p}{p_c}$	$\frac{\rho}{\rho_c}$	$\log \frac{p}{p_c}$	$\log \frac{\rho}{\rho_c}$	a (m/sec)	η^a ($\frac{g}{sec\,cm} \times 10^{-4}$)	$\frac{Re^b}{M}$
						Nitrogen Model I				
-39.2	580	2.21	1.41×10^{-3}	24.9	10.1	1.40	1.00	470	2.85	2.33×10^{7}
-35	543	1.75	1.19×10^{-3}	19.7	8.51	1.29	0.930	454	2.72	1.99×10^{7}
-30	499	1.29	9.57×10^{-4}	14.5	6.85	1.16	0.836	435	2.57	1.62×10^{7}
-25	455	0.930	7.56×10^{-4}	10.5	5.41	1.02	0.733	416	2.41	1.30×10^{7}
-20	411	0.651	5.86×10^{-4}	7.33	4.19	0.865	0.622	395	2.22	1.04×10^{7}
-15	367	0.434	4.36×10^{-4}	4.89	3.13	0.689	0.496	374	2.07	7.91×10^{6}
-10	323	0.277	3.17×10^{-4}	3.12	2.27	0.494	0.356	350	1.88	5.90×10^{6}
-5	279	0.163	2.16×10^{-4}	1.84	1.55	0.265	0.190	326	1.69	4.17×10^{6}
0	235	0.0888	1.40×10^{-4}	1	1	0	0	299	1.48	2.83×10^{6}
						Nitrogen Model II				
-13.1	350	0.367	3.88×10^{-4}	4.13	3.77	0.616	0.576	365	2.00	7.3×10^{6}
-10	323	0.276	3.16×10^{-4}	3.11	2.26	0.493	0.354	350	1.88	5.88×10^{6}
-8	305	0.225	2.73×10^{-4}	2.53	1.95	0.403	0.290	340	1.81	5.13×10^{6}
-5	279	0.164	2.17×10^{-4}	1.85	1.55	0.267	0.190	326	1.69	4.19×10^{6}
-3	261	0.129	1.83×10^{-4}	1.45	1.31	0.161	0.117	315	1.60	3.60×10^{6}
0	235	0.0888	1.40×10^{-4}	1	1	0	0	299	1.48	2.83×10^{6}

^{a}For nitrogen.

^{b}For $l = 1$ m.

Table 3B

N_2 MODELS I AND II

h (km)	T (°K)	p (atm)	(g/cm³)	$\frac{p}{p_c}$	$\frac{\rho}{\rho_c}$	$\log \frac{p}{p_c}$	$\log \frac{\rho}{\rho_c}$	a (m/sec)	η^a ($\frac{g}{sec\ cm} \times 10^{-4}$)	λ (cm)	$\frac{Re}{M}^b$
0	235	8.88×10^{-2}	1.40×10^{-4}	1	1	0	0	299	1.47	4.78×10^{-5}	2.85×10^{6}
20	220	5.30×10^{-3}	8.91×10^{-6}	5.97×10^{-2}	6.36×10^{-2}	-1.22	-1.20	289	1.40	7.39×10^{-4}	1.84×10^{5}
50	198	5.77×10^{-5}	1.08×10^{-7}	6.50×10^{-4}	7.71×10^{-4}	-3.19	-3.11	274	1.28	5.88×10^{-1}	2.31×10^{3}
60	190	9.95×10^{-6}	1.94×10^{-8}	1.12×10^{-4}	1.39×10^{-4}	-3.95	-3.86	269	1.24	3.23×10^{0}	4.21×10^{2}
80	234	5.25×10^{-7}	8.30×10^{-10}	5.91×10^{-6}	5.93×10^{-6}	-5.23	-5.23	298	1.47	8.08×10^{1}	1.68×10^{1}
100	278	4.61×10^{-8}	6.14×10^{-11}	5.19×10^{-7}	4.39×10^{-7}	-6.29	-6.36	325	1.68	1.14×10^{2}	1.19×10^{0}
150	389	3.71×10^{-10}	3.53×10^{-13}	4.18×10^{-9}	2.52×10^{-9}	-8.38	-8.60	385	2.19	2.19×10^{4}	6.21×10^{-2}
200	500	1.08×10^{-11}	8.00×10^{-15}	1.22×10^{-10}	5.71×10^{-11}	-9.91	-10.2	436	2.58	1.01×10^{6}	1.35×10^{-4}

^{a}For nitrogen.

^{b}For $\ell = 1$ m.

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